

Seasonal melting and the formation of sedimentary rocks on Mars, with predictions for the Gale Crater mound

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17 ABSTRACT

18 A model for the formation and distribution of sedimentary rocks on Mars
 19 is proposed. The rate-limiting step is supply of liquid water from seasonal
 20 melting of snow or ice. The model is run for a $O(10^2)$ mbar pure CO₂ atmo-
 21 sphere, dusty snow, and solar luminosity reduced by 23%. For these conditions
 22 snow only melts near the equator, and only when obliquity $\gtrsim 40^\circ$, eccentricity
 23 $\gtrsim 0.12$, and perihelion occurs near equinox. These requirements for melting are
 24 satisfied by 0.01–20% of the probability distribution of Mars’ past spin-orbit
 25 parameters. Total melt production is sufficient to account for aqueous alter-
 26 ation of the sedimentary rocks. The pattern of seasonal snowmelt is integrated
 27 over all spin-orbit parameters and compared to the observed distribution of
 28 sedimentary rocks. The global distribution of snowmelt has maxima in Valles
 29 Marineris, Meridiani Planum and Gale Crater. These correspond to maxima
 30 in the sedimentary-rock distribution. Higher pressures and especially higher
 31 temperatures lead to melting over a broader range of spin-orbit parameters.
 32 The pattern of sedimentary rocks on Mars is most consistent with a Mars pa-
 33 leoclimate that only rarely produced enough meltwater to precipitate aqueous
 34 cements and indurate sediment. The results suggest intermittency of snowmelt
 35 and long globally-dry intervals, unfavorable for past life on Mars. This model
 36 makes testable predictions for the Mars Science Laboratory rover at Gale
 37 Crater. Gale Crater is predicted to be a hemispheric maximum for snowmelt
 38 on Mars.

39 *Keywords:* MARS, CLIMATE; MARS, SURFACE; MARS, ATMOSPHERE;
 40 GEOLOGICAL PROCESSES; MARS

1 Introduction

Climate models struggle to maintain annual mean temperatures $\bar{T} \gtrsim 273\text{K}$ on Early Mars (Haberle, 1998) (Forget et al., submitted manuscript, 2012). Seasonal melting can occur for annual maximum temperatures $T_{max} \gtrsim 273\text{K}$, which is much easier to achieve. Therefore, seasonal melting of snow and ice is a candidate water source for surface runoff and aqueous mineralization on Mars. Surface temperatures $\sim 300\text{K}$ occur at low latitudes on today's Mars. However, seasonal melting of surface-covering, flat-lying snowpack does not occur because of (1) evaporative cooling and (2) cold-trapping of snow and ice near the poles or at depth. Reduced solar luminosity for Early Mars makes melting more difficult (Squyres and Kasting, 1994).

Toon et al. (1980) modelled the control of Milankovitch cycles on Mars ice temperatures. Melting is favored when snow is darkened by dust, and when evaporative cooling is reduced by increased pressure. Jakosky and Carr (1985) suggested that equatorial snowpacks would form at high obliquity. They pointed out that melt could contribute to the observed low-latitude erosion. Clow (1987) modeled snowmelt due to the solid-state greenhouse effect. He tracked meltwater migration to the base of the snowpack. Several authors have modeled melting on steep slopes as a candidate water source for young midlatitude gullies (Costard et al., 2002; Christensen, 2003). Hecht (2002) considered the energy balance for water at the melting point in gully alcoves. Williams et al. (2008, 2009) modeled melting of relatively clean snow overlain by a thin, dark lag deposit. They found melt rates $\sim 1 \text{ kg/m}^2/\text{hr}$ on steep slopes, but argue that this is sufficient to form gullies through either fluvial or debris-flow incision. Morgan et al. (2010) used a 1D atmospheric model to examine water ice melting and CO_2 frost accumulation. No author has studied Early Mars cold-traps, although Schorghofer's model (Schorghofer and Aharonson, 2005; Schorghofer, 2007b, 2010) has been used to track cold-traps for subsurface ice over the last 5 Ma (Schorghofer, 2007a).

The first purpose of this paper is to extend the global snowmelt models by integrating a new thermal model over all spin-orbit parameters, while accounting for cold-traps. Chaotic diffusion in the solar system makes it almost certain that Mars' obliquity (ϕ) has ranged twenty times more widely than Earth's obliquity over billion-year periods, and that Mars' eccentricity has had a long-term variance twice that of the Earth (Touma and Wisdom, 1993; Laskar and Robutel, 1993; Laskar et al., 2004; Laskar, 2008). These wide swings cause large variations in insolation and propensity to melt (Figure 1).

The second purpose of this paper is to understand the water source for sedimentary rock formation on Mars (Malin and Edgett, 2000). We focus on the hypothesis that supply of water from seasonal melting was the limiting

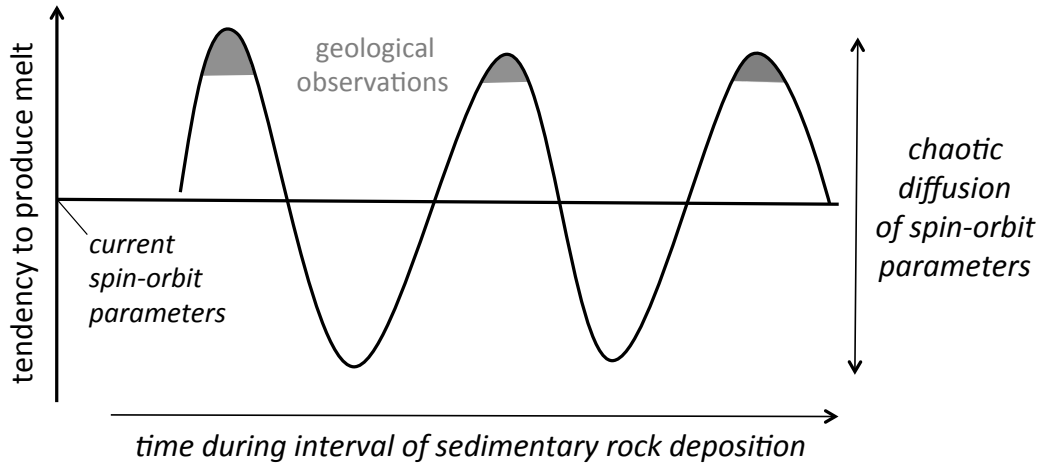


Fig. 1. Motivation for this paper. Mars underwent tens to thousands of spin-orbit oscillations during the interval of sedimentary-rock deposition. Three are shown schematically in this sketch. The geologic record of metastable surface liquid water is a wet-pass filter of Mars climate history. Mars orbital parameters vary over a wide range, resulting in a correspondingly wide range in tendency to melt. Therefore, evidence that the sedimentary rocks formed in a small fraction of Mars’ history (Lewis et al., 2010) suggests that negligible melting occurred under mean orbital forcing. If Mars sedimentary rock record only records orbital conditions that permitted surface liquid water, modeling average orbital conditions is neither sufficient nor appropriate. Instead, it is necessary to calculate snowmelt for the full range of orbital elements that Mars likely sampled over the time interval of sedimentary rock deposition. These predictions can then be compared to observations. Because of the evidence for orbital pacing of sedimentary rock accumulation (Lewis et al., 2008), transient warming events are not shown, but may have been critical for generating geomorphically effective runoff - see §8.4.

81 step in the formation of sedimentary rocks on Early Mars (e.g., Wendt et al.,
82 2011). Existing evidence for snowmelt-limited sedimentary rock formation is
83 discussed in §2.

84 If surface liquid water availability was necessary for sedimentary rock forma-
85 tion, then the spatial distributions of liquid water availability and sedimen-
86 tary rock detections should correspond. §3 analyzes the global sedimentary
87 rock distribution. In the only previous global model of sedimentary rock for-
88 mation on Mars, Andrews-Hanna et al. (2007) tracked groundwater flow in a
89 global aquifer that is recharged by a broad low-latitude belt of precipitation.
90 Groundwater upwelling is focussed in low-lying areas, generally consistent with
91 the observed distribution of sedimentary rocks (Andrews-Hanna et al., 2010;
92 Andrews-Hanna and Lewis, 2011). This model assumes $\bar{T} > 273\text{K}$, in order to
93 avoid the development of an impermeable cryosphere. Especially in light of
94 the Faint Young Sun predicted by standard solar models, temperatures this
95 high may be unsustainable for the long periods of time required to form the

sedimentary rocks (Haberle, 1998; Tian et al., 2010) (Forget et al., submitted manuscript, 2012). The model described (§4) and analyzed (§5) in this paper assumes that liquid water is supplied from locally-derived snowmelt, rather than a deep global aquifer. Groundwater flow is confined to shallow, local aquifers perched above the cryosphere. Annually- and planet-averaged temperatures remain similar to today’s, which reduces the required change in climate forcing from the present state. If Mars climate once sustained $\bar{T} > 273\text{K}$, then it must have passed through climate conditions amenable to snowmelt en route to the modern desert (McKay and Davis, 1991). The converse is not true.

Model predictions for different paleoclimate parameters are compared to global data in §6. §7 makes testable predictions for Gale Crater, the target for the Mars Science Laboratory (MSL) mission (Milliken et al., 2010).

The discussion (§8) includes comparison of the snowmelt model to alternative hypotheses such as global groundwater (Andrews-Hanna et al., 2007, 2010; Andrews-Hanna and Lewis, 2011) and cryogenic weathering within ice sheets (Catling et al., 2006; Niles and Michalski, 2009). The conclusions of this study are stated in §9.

This scope of this paper is forward modeling of snowmelt production as a function of (unknown) Early Mars climate parameters. Beyond a qualitative discussion in §7 and §8, there is no attempt to physically model the processes running from snowmelt production to sedimentary rock formation. A computationally inexpensive 1D model allows us to sweep over a large parameter space. The trade-off is that 1D models cannot track the effect of topographically-forced planetary waves on the atmospheric transport of water vapor, which controls snow precipitation (Vincendon et al., 2010; Forget et al., 2006; Colaprete et al., 2005). Any 1D snow location prescription is therefore an idealization.

2 Snowmelt hypothesis

Liquid water is required to explain sedimentary rock texture and bulk geochemistry along the Mars Exploration Rover *Opportunity* traverse, and there is strong evidence for extending this conclusion to other light-toned, sulfate-bearing sedimentary rocks on Mars (Bibring et al., 2007; McLennan and Grotzinger, 2008; Roach et al., 2010; Murchie et al., 2009a; Weitz et al., 2008). The hypothesis in this paper is that the water source for sedimentary rocks on Early Mars was seasonal melting, and that liquid water was infrequently available so that melt availability was the limiting factor in forming sedimentary rocks. “Sedimentary rocks” is used to mean units comprised of chemical

precipitates or siliciclastic material cemented by chemical precipitates, usually sulfates. These are recognized from orbit as light-toned layered sedimentary deposits (Malin et al., 2010) that characteristically show diagnostic sulfate features in the near-infrared. This definition excludes layered phyllosilicates, which usually predate sulfates (Bibring et al., 2006; Ehlmann et al., 2011).

2.1 *What is the evidence that sediment lithification on Mars requires liquid water?*

Erosion to form cliffs and boulders (Malin and Edgett, 2000), ejection of meter-size boulders from small, fresh craters (Golombek et al., 2010), resistance to crushing by rover wheels, and microscopic texture (Okubo, 2007) show that most light-toned sedimentary deposits (hereafter “sedimentary rocks”) are indurated or lithified. Lithification involves compaction and cementation. Water is required to form aqueous cements, and for fluvial sediment transport. At the *Opportunity* landing site, evaporitic sandstones (60% chemical precipitates by weight on an anhydrous basis) record groundwater recharge and aqueous cementation, surface runoff, and shallow lithification (McLennan et al., 2005; McLennan and Grotzinger, 2008). Aqueous minerals are present in sedimentary rocks throughout Meridiani and the Valles Marineris (Bibring et al., 2007). Murchie et al. (2009a) argue for water-limited lithification of the Valles Marineris sedimentary rocks. Subsurface pressure-solution recrystallisation can occlude porosity and lithify weak evaporites at ~ 30 bars, without aqueous cementation (Warren, 2006). But in this case water is probably still needed to form evaporites at 60% by weight. Some layered sedimentary deposits on Mars might not require liquid water to form, but these deposits are usually younger or at higher latitudes than the sulfate-bearing layered sedimentary rocks (Hynek et al., 2003; Bridges et al., 2010; Fenton and Hayward, 2010).

Outcrop thermal inertia (TI) is almost always low at 100m scale, so these rocks were either never strongly cemented or have been weakened after exhumation (Edwards et al., 2009).

2.2 *When did sulfate-bearing sedimentary rocks form?*

Sulfate-bearing sedimentary rocks occur relatively late in the stratigraphic sequence of evidence for stable surface liquid water on Mars (Murchie et al., 2009b; Fassett and Head, 2008; Massé et al., 2012; Mangold et al., 2010; Thollot et al., 2012). Formation of sedimentary rocks peaked in the Hesperian (Carr and Head, 2010), well after the peak of phyllosilicate formation on Mars (Ehlmann et al., 2011; Fassett and Head, 2011; Salvatore et al., 2010;

171 Bibring et al., 2006). Phyllosilicates are sometimes interbedded with older
 172 sulfate-bearing sedimentary rocks (Ehlmann et al., 2011), but those phyllosil-
 173 icates may be reworked (Barnhart and Nimmo, 2011). Sedimentary rocks also
 174 postdate almost all of the large-scale, regionally integrated highland valley net-
 175 works of the Late Noachian/Early Hesperian (Carr and Head, 2010; Fassett
 176 and Head, 2011), and are spatially separated from these “classic” valley net-
 177 works (Hynek et al., 2010). Therefore, the observed sedimentary rocks cannot
 178 be the terminal deposits of the classic valley networks. The climate that cre-
 179 ated the classic valley networks could have been different from the climate that
 180 formed the sedimentary rocks (Andrews-Hanna and Lewis, 2011). Sedimen-
 181 tary rocks do contain some channels, often preserved in inverted relief (Edgett,
 182 2005; Burr et al., 2009, 2010). Finally, many sedimentary rocks postdate the
 183 large impacts of the Late Heavy Bombardment, and many have quasi-periodic
 184 bedding suggesting orbitally-paced deposition (Lewis et al., 2008, 2010). These
 185 observations are inconsistent with brief bursts of rapid sedimentary rock for-
 186 mation during impact-induced greenhouse events.

187 2.3 What existing data supports the snowmelt hypothesis?

188 Liquid water was in short supply even at the time of sedimentary rock for-
 189 mation at the *Opportunity* landing site. Mineralogy indicates low water/rock
 190 ratios during alteration, and that the cumulative duration of water-rock in-
 191 teraction was ≤ 100 ka (Berger et al., 2009; Hurowitz and McLennan, 2007;
 192 Elwood Madden et al., 2009). Weathering at Meridiani Planum was either iso-
 193 chemical or at low water/rock ratio or both (Ming et al., 2008), consistent with
 194 a rare trickle of snowmelt. Low specific grind energy of sandstones indicates
 195 weak aqueous cementation (Herkenhoff et al., 2008). The present day extent
 196 of sedimentary rock outcrops is small, and the persistence of opal, jarosite
 197 and olivine in rocks (and olivine in soils) indicates minimal water-rock inter-
 198 action since those minerals crystallized (Tosca and Knoll, 2009; Olsen and
 199 Rimstidt, 2007; Yen et al., 2005). Away from the sedimentary rocks them-
 200 selves, aqueous mineralization was minor or absent elsewhere on the planet at
 201 the time when most sulfate-bearing sedimentary rocks formed (Murchie et al.,
 202 2009b; Salvatore et al., 2010; Mustard et al., 2009; Fassett and Head, 2011;
 203 Hausrath et al., 2008). Globally, soils formed “with little aqueous alteration
 204 under conditions similar to those of the current Martian climate” (Bandfield
 205 et al., 2011), and elemental profiles indicate top-down mobilization of soluble
 206 elements (Amundson et al., 2008; Arvidson et al., 2010).

207 Dividing the total thickness of sedimentary rock deposits by the thickness of
 208 quasi-periodic layers and then multiplying by the obliquity periods thought
 209 to pace accumulation suggests that the sedimentary rocks formed in 1-10
 210 Ma (Lewis et al., 2010). This is a small fraction of Mars’ history. Geomor-

phic evidence that the Mars surface environment has only marginally supported surface liquid water since the Noachian includes mean erosion rates ~ 1 atom/year (Golombek et al., 2006), together with a sharp post-Noachian decline in valley network formation and crater infilling (Fassett and Head, 2008; Forsberg-Taylor et al., 2004). These data argue for a short-lived and downward-infiltrating post-Noachian water supply, suggestive of transient liquid water that is generated only during brief melt events.

2.4 *How could brief pulses of snowmelt form kilometer-thick accumulations of sedimentary rock?*

Antarctica's McMurdo Dry Valleys are a terrestrial analog for seasonal-melt-limited fluvial erosion and sedimentary rock formation at $\bar{T} < 273\text{K}$ (Doran et al., 2010; Lee and McKay, 2003; Marchant and Head, 2007). Weathering and mineralization is confined to lakes, hyporheic zones, and a shallow active layer. However, seasonal river discharges reach $20 \text{ m}^3 \text{ s}^{-1}$ (McKnight, 2011), fluvial valleys incise $> 3\text{m}$ deep into granite (Shaw and Healy, 1980), and annually-averaged weathering intensity within the hyporheic zone is greater than in temperate latitudes (Nezat et al., 2001). Ions are concentrated within ice-covered lakes by sublimation. Outcrops of gypsum, carbonate evaporites, and algal limestone sediments show that sediments have accumulated at the base of melt-fed perennial lakes for 300,000 years (McKay et al., 1985; Hendy, 2000). Dry Valley Drilling Project cores show lithification in older horizons (McKelvey, 1981).

Order-of-magnitude energy and mass balance shows that brief, rare pulses of snowmelt provide enough water to form the kilometers of sedimentary rock observed on Mars. For solar luminosity reduced by 23%, peak noontime insolation at Mars at perihelion on a moderate-eccentricity ($e=0.15$) orbit is $\approx 630 \text{ W/m}^2$. If the snowpack is dusty then its albedo will be that of Mars dust, 0.28 (Putzig et al., 2005). During melting, radiative losses are $\sigma T_{\text{melt}}^4 \approx 320 \text{ W/m}^2$, and for a 200 mbar atmosphere a reasonable value for wind-speed dependent sublimation losses into dry air is $\sim 60 \text{ W/m}^2$. Conductive losses will be roughly one-half the diurnal temperature range divided by the diurnal skin depth, giving 60 W/m^2 for the snowpack material properties in Carr and Head (2003) and a 100K diurnal cycle of surface temperature. Greenhouse forcing from a 200 mbar CO_2 atmosphere equilibrated with a 230K daily mean surface temperature is $\sim 60 \text{ W/m}^2$ (from detailed radiative transfer calculations: Appendix B). Neglecting all other gains and losses, the net energy available for melting is therefore $630(1-0.28) - 320 - 60 - 60 + 60 \sim 100 \text{ W/m}^2$, equivalent to approximately $1 \text{ kg/m}^2/\text{hr}$ snowmelt. The total water required to form the 800m-thick Meridiani sediments depends on the water/rock mass ratio (W/R) during alteration. W/R is given as $\lesssim 1$ by Berger et al. (2009) and $\lesssim 300$ by

251 Hurowitz and McLennan (2007). This corresponds to a time-integrated melt
 252 column of either $\lesssim 0.3$ km or $\lesssim 100$ km, respectively, for a bulk Meridiani sand-
 253 stone density of 2.3, and ignoring the contribution of water bound in hydrated
 254 minerals to the solid mass of the deposit. Assuming melting occurs for 10% of
 255 each sol, the melt season lasts for 10% of the year, and 10% of melt is avail-
 256 able for alteration, snowmelt production for only 30 Kyr (for $W/R = 1$) or 10
 257 Myr (for $W/R = 300$) provides enough snowmelt to reach the upper limit on
 258 W/R for the entire Meridiani sandstone. The period in question lasted $O(10^9)$
 259 years, so climate and orbital conditions favorable for surface liquid water at
 260 Meridiani are only needed for $<1\%$ of the time. This is consistent with the
 261 wet-pass filter sketched in Figure 1.

262 **3 Distribution of sedimentary rocks on Mars**

263 The Mars Orbiter Camera Narrow Angle (MOC NA) team documented $\sim 4,000$
 264 “layered rock outcrops of probable or likely sedimentary origin” (Malin and
 265 Edgett, 2000, 2001; Malin et al., 2010). Details of our analysis of these data
 266 are give in Appendix A. The resulting distribution of sedimentary rocks on
 267 Mars (Figure 2) suggests that surface water availability was narrowly concen-
 268 trated near the equator and at low elevations. 64% of sedimentary rocks are
 269 within 10° of the equator, 60% when the Valles Marineris region is excluded
 270 (Figure 2a). Blanketing by young mantling deposits may contribute to the
 271 paucity of sedimentary rocks poleward of 35° , but cannot explain the rarity of
 272 sedimentary rocks at 10 - 35° latitude relative to the equatorial belt. The $\pm 10^\circ$
 273 band is not unusual in thermal inertia, dust cover index, albedo, or surface age
 274 distribution, so a dependence of sedimentary rock on these parameters could
 275 not explain the latitudinal distribution.

276 On average, sedimentary rocks are lower than ancient terrain by 2km (Figure
 277 2b). On Earth sedimentary rocks are low-lying because of sediment transport
 278 by regional-integrated channel networks. Evidence for regionally-integrated
 279 channel networks on Mars mostly predates the sedimentary rock era (Carr and
 280 Head, 2010; Fassett and Head, 2011). The low-elevation bias is independent of
 281 the equatorial concentration. Therefore, the low-elevation bias is reflective of
 282 a planetwide, non-fluvial process that occurs preferentially at low elevations.

283 Sedimentary rock abundance does not decline monotonically away from the
 284 equator. Abundance away from the equator is much less than in the equatorial
 285 sedimentary-rock belt, but “wings” of increased sedimentary rock abundance
 286 are found nearly symmetric about the equator at 25 - 30° S and 20 - 30° N. The
 287 sedimentary rocks in the southern wing are regionally associated with clusters
 288 of large alluvial fans (Figure 2d).

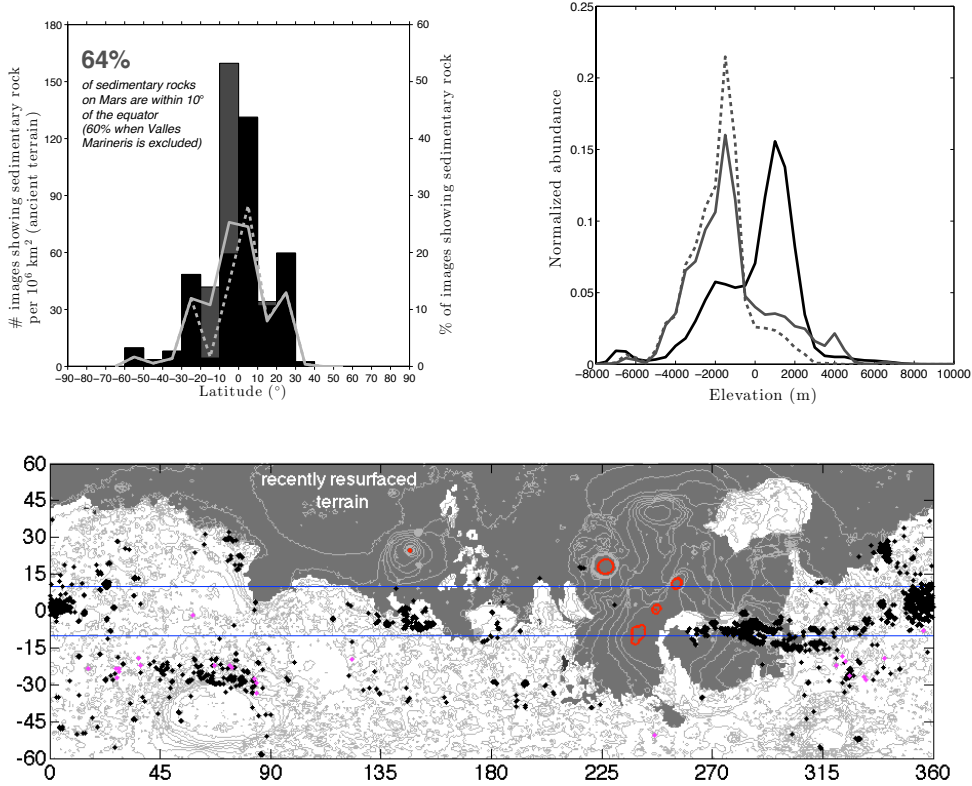


Fig. 2. **(a)** Latitudinal dependence of sedimentary rocks, masking out recently resurfaced terrain (Appendix A). Latitude bin size is 10°. Histogram corresponds to number of images (left axis). Dark gray bars are the contribution from the Valles Marineris region, and black bars represent the rest of the planet. Lines correspond to the percentage of images showing sedimentary rock (right axis). Dashed line is the percentage of images showing sedimentary rocks once the Valles Marineris region is excluded. **(b)** Elevation dependence of sedimentary rocks, masking out recently resurfaced terrain. Elevation bin size is 500m. Gray line is normalized histogram of terrain with sedimentary rocks, and black line is histogram of all ancient terrain. Dotted gray line is the normalized histogram of terrain with sedimentary rocks, after masking out Valles Marineris. Median sedimentary rock elevation is ~ 2 km lower than median ancient terrain. **(c)** Distribution of sedimentary rocks on Mars (black dots, from Malin et al. (2010)). Alluvial fans are also shown (purple dots, from Kraal et al. (2008)). Blue horizontal lines highlight the $\pm 10^\circ$ latitude band. Dark gray shading corresponds to recently resurfaced terrain. Light gray contours show topography, with the +10km contour shown in red.

289 4 Model

290 This section describes the seasonal melt model. §4.1 describes the model frame-
 291 work and assumptions. §4.2 describes the 1D snowpack thermal model. §4.3
 292 describes the potential-well approximation for warm-season snow locations,
 293 and §4.4 explains how results from 1D models are combined to produce pre-
 294 dictions.

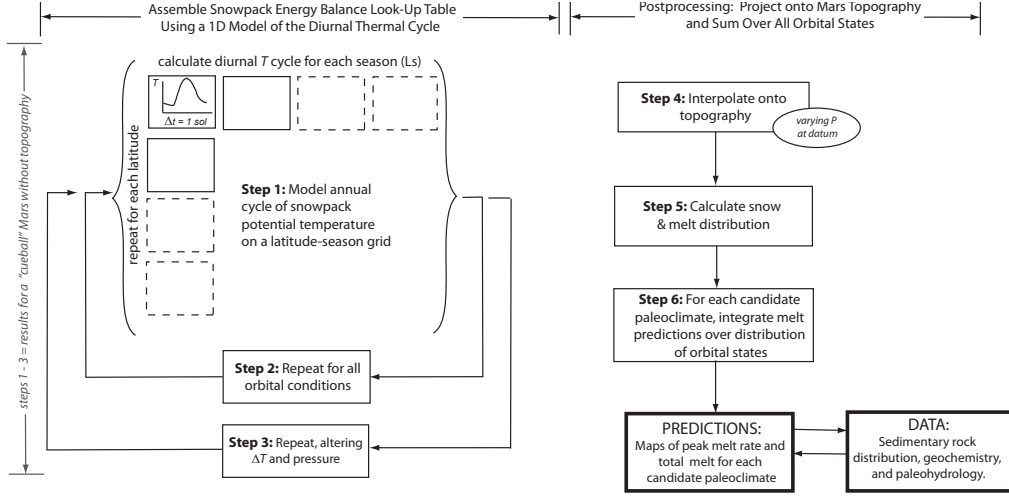


Fig. 3. Workflow of the Early Mars seasonal melting model. See text for details.

4.1 Overview of model framework

Controls on Mars snowmelt include:–

- *Spin-orbit properties* $\mathbf{O} = \{\phi, e, L_p, L_s, \text{latitude}\}$ control the distribution of sunlight at a given location. These include obliquity ϕ , eccentricity e , solar longitude of perihelion L_p , solar longitude L_s , and latitude. Milankovitch parameters $\mathbf{O}' = \{\phi, e, L_p\}$ oscillate or circulate on 10^4 – 10^6 yr frequencies, and ϕ shows chaotic shifts at ~ 250 Myr intervals (Head, 2011). We iterate over all the spin-orbit properties that have probably been encountered by Mars over the last 3.5 Ga (Steps 1–2 in Figure 3).
- *Climate parameters* $\mathbf{C} = \{P, \Delta T, f_{\text{snow}}\}$ include atmospheric pressure P (assumed to be mostly CO_2), freezing-point depression/non- CO_2 greenhouse forcing ΔT , and snow coverage fraction f_{snow} . These are iterated over a large range (Step 3 in Figure 3). They are assumed to vary slowly relative to changes in spin-orbit properties.
- *Surface material properties, insolation.* These are held fixed for the ensemble of model runs. Results are sensitive to snowpack TI and albedo. Parameter choices and sensitivity tests are discussed in Appendix D.

The predictions of the snowmelt hypothesis (§2.6) are evaluated as follows (Figure 3). First, for a given trial set of past climate parameters \mathbf{C} and orbital parameters \mathbf{O} , snow temperature for all seasons and latitudes is calculated using a 1D surface energy balance model (Step 1 in Figure 3). Using the 1D model output for a range of P (Step 3), the potential annual-average snow sublimation rates are mapped onto topography (Step 4). Warm-season snow is assigned to locations most favorable for interannually persistent snow (Figure 10). Snowmelt occurs when temperatures at these locations exceed freezing (Step 5). These results provide latitude-season diagrams that are appropriate

for a flat Mars. Results for this fictitious “cueball” Mars are analyzed in §5.3. To map the results onto real Mars topography, a sequence of latitude-season output for different P but the same \mathbf{O}' is stacked in elevation (results given in §5.4). The output at this stage consists of maps of snow stability for the given \mathbf{C} and \mathbf{O}' , along with time series of snow temperature and melt rates. Next, the framework loops over all possible Early Mars \mathbf{O} (Step 6), convolving the outputs with the \mathbf{O}' probability distribution function (Laskar et al., 2004). The output is now a map of predicted snowmelt on Mars integrated over geologic time, and this can be compared to observed sedimentary rock abundance and thickness data (§5.5). These maps are computed for many plausible \mathbf{C} . Assuming the snowmelt model is correct, the \mathbf{C} that produces the best correlation between model predictions and data is the best-fit Early Mars climate (§6.1). Interpolating in the melt rate output gives a predicted time series at Gale Crater (§7).

4.2 Thermal model

Surface liquid water is always unstable to evaporation on a desert planet (Richardson and Soto, 2008a,b). However, transient liquid water can occur metastably if temperatures exceed the freezing point, and if P exceeds the triple point (in order to prevent internal boiling) (Hecht, 2002).

These conditions are modeled using a 1D thermal model (Figure 4). When temperature exceeds $(273.15\text{K} - \Delta T)$, melting occurs and buffers the temperature at the melting point. Within the snowpack, material properties are assumed uniform with depth and heat flow is by conduction and solar absorption only. When melt is not present, energy balance in the subsurface layer adjacent to the surface is given for unit surface area by (Figure 4)

$$\rho C_s \Delta z \frac{\partial T_1}{\partial t} = \frac{k}{\Delta z} \frac{\partial^2 T}{\partial z^2} - \underbrace{\epsilon \sigma T^4 + LW\downarrow + (1 - s_r)Q(1)SW\downarrow}_{\text{radiative terms}} - \underbrace{S_{fr} - L_{fr}}_{\text{free convection}} - \underbrace{S_{fo} - L_{fo}}_{\text{forced convection}} \quad (1)$$

and energy balance at depth z within the snowpack is given by (Figure 4)

$$\rho C_s \Delta z \frac{\partial T_K}{\partial t} = \frac{k}{\Delta z} \frac{\partial^2 T}{\partial z^2} + \underbrace{(1 - s_r)Q(z)SW\downarrow}_{\text{solid-state greenhouse}} \quad (2)$$

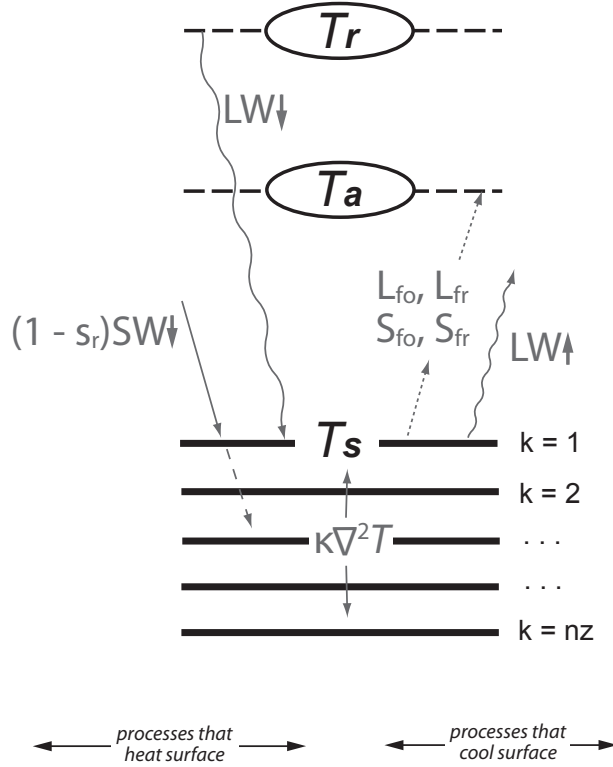


Fig. 4. Vertical discretization and energy flow in the 1D model. Solid horizontal lines correspond to solid surface layers numbered $K = \{1 \dots nz\}$, dashed horizontal lines correspond to atmospheric layers. T_r is the effective atmospheric radiative temperature, T_a is the atmospheric surface layer temperature, and T_s is the ground surface temperature. The diagonal arrows correspond to energy fluxes: $LW \downarrow$ for greenhouse effect, $(1 - s_r)SW \downarrow$ for insolation attenuated by Rayleigh scattering, $LW \uparrow$ for backradiation, and $\{L_{fo}, L_{fr}, S_{fo}$ and $S_{fr}\}$ for the turbulent fluxes. Some insolation penetrates into the snowpack (dashed continuation of insolation arrow). $\kappa \nabla^2 T$ corresponds to conductive diffusion.

Here, ρ is snow density, C_s is snow specific heat capacity, Δz is the thickness of the subsurface layer whose upper boundary is the surface, T is the temperature at subsurface level $K = \{1, 2, \dots, nz\}$ (Figure 4), k is snow thermal conductivity, ϵ is the longwave emissivity of ice, $LW \downarrow$ is downwelling long-wave radiation, s_r is the Rayleigh-scattering correction factor, $Q_{\{1,2,\dots,nz\}}$ is the fraction of sunlight absorbed at level z (Appendix C), $SW \downarrow$ is solar flux per unit surface area, S_{fr} corresponds to free sensible heat losses driven by atmosphere-surface temperature differences, L_{fr} corresponds to evaporative cooling by free convection when the atmosphere has relative humidity < 1 , S_{fo} corresponds to forced sensible heat losses caused by cool breezes over warm ground, and L_{fo} corresponds to additional evaporative cooling when the wind is nonzero. Snow albedo, α , is $1 - \int Q_z dz$. All results presented in this paper are for the 3.5 Gya solar luminosity reported by Bahcall et al. (2001), 23% below present.

361 The 1D model draws on previous work by Toon et al. (1980); Clow (1987);
 362 Richardson and Mischna (2005); Williams et al. (2008); Dundas and Byrne
 363 (2010); and Liston and Winther (2005). A simpler version is discussed in Kite
 364 et al. (2011b,c). Representative output is shown in Figure 5. Details of the flux
 365 parameterizations, melt handling, and run conditions are given in Appendix
 366 B.

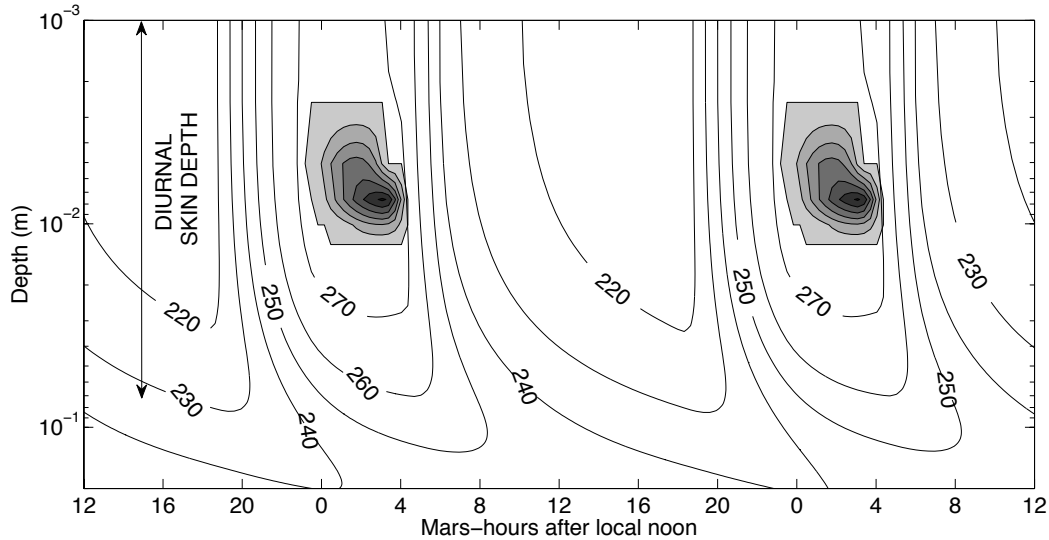


Fig. 5. Daily cycle of temperature and melting in the upper 15cm of snowpack. Grayscale corresponds to melt fraction at a constant temperature of 273.15K, with contours of 0.1 (edge of gray region corresponds to zero melt). Temperature contours are shown wherever the snow is not melting, at intervals of 10K. The blocky outline of the “melt = 0” contour corresponds to the 1200s resolution at which model output is stored for subsequent processing: the underlying timestep is 12s. $L_s = 0^\circ$, $L_p = 0^\circ$, $\phi = 50^\circ$.

367 4.3 Snow location prescription: the potential-well approximation

368 For ~ 100 mbar CO_2 , the model predicts melt only during the warmest season,
 369 and usually within a diurnal skin depth of the surface. Experiments, observa-
 370 tion and theory agree that warm-season snow and ice within this depth range
 371 on Mars is in cold-trap equilibrium with orbital forcing (Mellon and Jakosky,
 372 1995; Hudson and Aharonson, 2008; Hudson et al., 2009; Schorghofer and
 373 Aharonson, 2005; Boynton et al., 2002). Because we are interested in above-
 374 freezing T only where snow is present, the quantity of interest is the annual-
 375 maximum T experienced by the cold traps, whose location depends on orbital
 376 conditions and topography. For most orbital conditions, this T is below freez-
 377 ing, so the greatest interest is in the orbital conditions that maximize the
 378 cold-traps’ annual maximum T .

379 In the case of no topography (cueball planet), the location of the cold traps

depends on orbital conditions only. For this case we find the location of cold traps in the following manner: for each \mathbf{O}' , the output of the thermal model for all seasons (L_s) and geographic locations \mathbf{x} is used to determine the \mathbf{x} where snow is most likely to be present during the melt season. Melt-season snow is assumed to be only found at locations where the annually-averaged sublimation is minimized (Figure 6). Converged thermal-model output for L_s is linearly interpolated in time using Kepler’s equation (Figure 6). All \mathbf{x} are then assigned an area-weighted rank, f , scaled from 0% (global minimum in annually-averaged sublimation; most favorable for snow accumulation) to 100% (global maximum in annually-averaged sublimation; least favorable to snow accumulation). Ice lost by melting is assumed to be recovered by refreezing close ($<100\text{km}$) to source. Warm-season snow is assumed not to occur above a critical f , termed f_{snow} (the percentage of the planet’s surface area that has warm-season snow). Using the $f(\mathbf{x})$ and f_{snow} , warm-season snow is assigned to favored geographic locations (Figure 7). In general, the critical f for warm-season snow will be greater than the critical f for interannually-persistent snow, so melting does not have to be supraglacial. The most favorable circumstances for aqueous alteration may be where melt occurs during a seasonal accumulation-ablation cycle which leaves bare soil during part of the year.

Melting is almost certain to occur when orbital forcing leads to annual-maximum temperatures above freezing at all latitudes (Figure 7). Thermal barriers $>10\text{ cm}$ thick can insulate snow against diurnal melting, but a sublimation lag covering all ice is logically impossible, and a debris lag covering all ice is unlikely. The albedo of pure, fresh, fine-grained snow is high enough to prevent melting, but contamination with dust is very likely (Appendix C). Twice-yearly transfer of the water ice reservoir across the equator to the cold high-obliquity winter pole would require unreasonably high seasonally reversing mean wind speeds.

With MOLA topography, annual-average sublimation rates are calculated as for the cueball case, but now for a range of P that spans mountaintop pressures and canyon pressures. The latitude- P grid is then interpolated onto latitude and longitude using MOLA topography (Figure 12). This assumes that Mars’ long-wavelength bedrock topography was in place before the sedimentary rock era, in agreement with geodynamic analysis (e.g., Phillips et al., 2001). This also neglects the effect of the adiabatic lapse rate on surface T , which is an acceptable approximation at the P of interest here (§8.1).

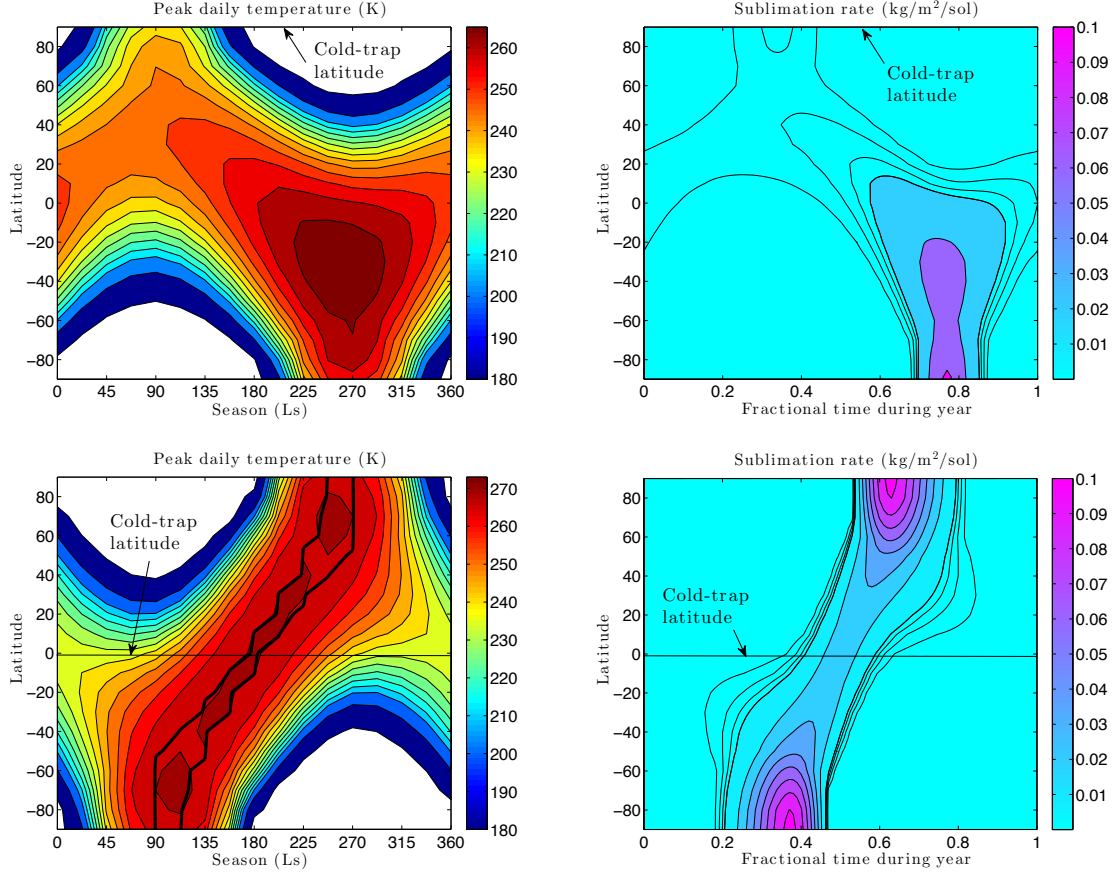


Fig. 6. Seasonal cycle of diurnal-peak temperature and diurnal-mean free sublimation rate for 3.5 Gya insolation, assuming flat topography. 145 mbar pure CO₂ atmosphere. **(a,b)** present-day orbital forcing ($\phi = 25.2^\circ$, $e = 0.093$, $L_p = 251^\circ$); **(c,d)** optimal conditions for melting – high- ϕ , moderate e , and L_p aligned with equinox. Contours of daily maximum surface temperature are drawn at 180K, 200K, and 210K and then at intervals of 5K up to a maximum of 270K, only reached in **(c)**. White shading corresponds to CO₂ condensation at the surface. Sublimation-rate contours are drawn intervals of 0.025 kg/m²/sol from 0 to 0.1 kg/m²/sol and then at intervals of 0.2 kg/m²/sol. At low e and low ϕ (similar to today, **(a,b)**), ice is stable at the poles, where temperatures never exceed freezing. In **(c,d)**, ice is most stable at the equator, and annual peak temperature exceeds freezing everywhere at some point during the year. The thick black line in **(c)** corresponds to subsurface melting at some point during the day. (No melting is predicted for modern orbital conditions.) The blockiness of this line corresponds to the underlying seasonal resolution (22.5° in L_s). Solid-state greenhouse effect warms the subsurface relative to the surface by up to several K.

4.4 Integrating over all orbital states

For each \mathbf{O}' , f is obtained for each \mathbf{x} , together with the potential snowpack temperatures and potential melt rates. Then, f_{snow} maps out the snow distri-

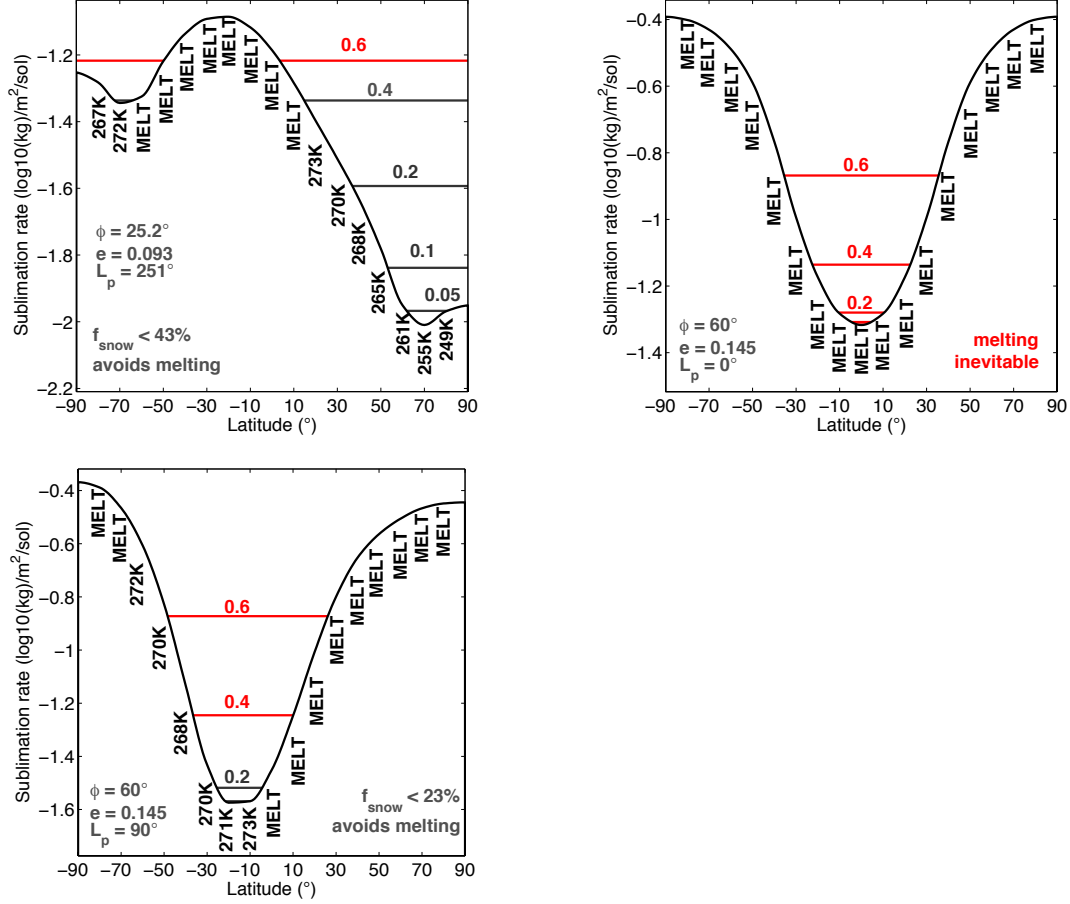


Fig. 7. To illustrate the potential-well approximation for finding warm-season snow locations. Nominal parameters (Table 1), 146 mbar atmosphere, flat topography. Curve corresponds to potential sublimation during a year. Temperatures are annual maxima. “MELT” denotes melting of snowpack at some point during the year, if it exists at that latitude. Horizontal lines correspond to f_{snow} values, assuming warm-season snow is found at locations that minimize annually-averaged sublimation, and colored red for values of f_{snow} that lead to melting. (a) Current orbital conditions. Massive ice or buried ice may exist in the southern hemisphere, but snowpack that persists through the warm-season is only likely in the far north, where temperatures are always below freezing. (b) Optimal orbital conditions. Melting occurs at all latitudes so melting is inevitable. (c) As for optimal orbital conditions, but for perihelion aligned with northern summer solstice. The short, intense northern summer displaces the potential-sublimation minimum to 20S. $f_{snow} > 23\%$ is needed for melting to occur under these circumstances. The latitude of first melting will be near the equator. Note that ‘273K’ is just below the melting point.

420 bution, and f_{snow} and ΔT map out the melt distribution. The resulting maps
 421 of snow and snowmelt are weighted using the \mathbf{O}' probability distribution func-
 422 tion (Laskar et al., 2004), and the weighted maps are summed. (Note that this
 423 is not the same as the median melt column from a large ensemble of solar

424 system integrations; Laskar et al. (2004).) Melt likelihood is then given by

$$M_{\mathbf{x}} = \int (T_{max,\mathbf{x}} > (273.15 - \Delta T)) (f_{snow} > f_{\mathbf{x}}) p(\mathbf{O}') d\mathbf{O}' \quad (3)$$

425 where the “greater than” operator returns 1 if true and 0 if false, and $p()$ is
426 probability.

Symbol	Parameter	Value and units	Source / rationale
Fixed parameters:			
A_{vonk}	von Karman's constant	0.4	
C_p	Specific heat, Mars air	770 J/kg/K	
C_s	Specific heat capacity of snow	1751 J/kg/K	Carr and Head (2003)
D_{air}	Mechanical diffusivity of air	$14 \times 10^{-4} \text{ m}^2/\text{s}$	Hecht (2002)
g	Mars surface gravity	3.7 m/s^2	
k_b	Boltzmann's constant	$1.38 \times 10^{-23} \text{ m}^2 \text{ kg s}^{-2}$	
k_{snow}	Thermal conductivity of snowpack	0.125 W/m/K	Carr and Head (2003)
m_c	Molar mass of CO ₂	0.044 kg	
m_w	Molar mass of H ₂ O	0.018 kg	
M_w	Molecular mass of H ₂ O	$2.99 \times 10^{26} \text{ kg}$	
$P_{atm,0}$	Pressure of atmosphere now	610 Pa	NSSDC
L_e	Latent heat of water ice sublimation	$2.83 \times 10^6 \text{ J/kg}$	Hecht (2002)
L_e	Latent heat of water ice melting	$3.34 \times 10^5 \text{ J/kg}$	Hecht (2002)
r_h	Relative humidity	0.25	
R_{gas}	Gas constant	8.3144 J/(mol K)	
$u_{s,ref}$	Reference near-surface wind speed	3.37 m/s	Millour et al. (2008) "MY24" average
z_o	Roughness length	0.1 mm	Polar snow. Brock et al. (2006)
z_{anem}	Anemometer height	5.53 m	Millour et al. (2008)
α	Albedo	≈ 0.28	Albedo of dusty snow. Appendix D.
ϵ	Emissivity of ice at thermal wavelengths	0.98	
ν_{air}	Kinematic viscosity of air	$6.93 \times 10^{-4} \text{ m}^2/\text{s}$	Hecht (2002)
ρ	Density of snowpack	350 kg m^{-3}	Carr and Head (2003)
ρ_0	Density of atmosphere (now)	0.02 kg m^{-3}	NSSDC
σ	Stefan-Boltzman constant	$5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$	
τ	Time of interest	3.5 Gyr ago	Murchie et al. (2009b)
	Mars semimajor axis	1.52366 AU	NSSDC
	Duration of 1 Mars sol	88775 s	
	Dust concentration	$\sim 2\%$ by volume	Appendix C
	Dust radius	4 μm	Appendix C
	Ice grain radius	1 mm	Appendix C
	Solar constant (now)	$1.361 \times 10^3 \text{ W/m}^2$	Kopp and Lean (2011)
Selected variables:			
O	Spin-orbit properties		
O'	Milankovitch parameters		
C	Climate parameters		
ϕ	Obliquity	0-80°	
b_{DB}	Dundas-Byrne "b"	f(P) (Appendix B)	Extrapolation from GCM runs
e	Eccentricity	0.00 – 0.16	
L_p	(Solar) longitude of perihelion	0 – 360°	
L_s	Solar longitude	0 – 360°	
M	Mean anomaly	0 – 360°	
P	Atmospheric pressure	24-293 mbar	
P_o	Atmospheric pressure at zero elevation	24-293 mbar	
ΔT	Non-CO ₂ greenhouse forcing	0-15 K	
f_{snow}	Fraction of planet surface area with warm-season snow	0-50%	
Q_k	Fraction of incident sunlight absorbed at level k	0-100%	
$LW\downarrow$	Greenhouse forcing		
$LW\uparrow$	Thermal emission by surface		
$SW\downarrow$	Insolation		
s_r	Rayleigh scattering correction factor		
L_{fo}	Latent heat losses by forced convection		
L_{fr}	Latent heat losses by free convection		
S_{fo}	Sensible heat lost by forced convection		
S_{fr}	Sensible heat lost by free convection		

Table 1

Selected parameters and variables.

428 5 Results and analysis

429 5.1 Controls on the occurrence of near-surface liquid water on Early Mars

430 Warm-season snow locations depend on sublimation rates. Diurnal-mean sub-
 431 limination is shown as a function of P in Figure 8. Losses due to free convection
 432 decrease with increasing P , because the greater atmospheric density dilutes
 433 the buoyancy of moist air (Appendix B). Losses due to forced convection in-
 434 crease with atmospheric density (Appendix B). Surface temperature increases
 435 monotonically with increasing greenhouse forcing, leading to an uptick in sub-
 436 limination rate for $P > 100$ mbar. Snow is most stable against sublimation when
 437 $P \sim 100$ mbar. Therefore, when the atmospheric pressure at zero elevation
 438 (P_o) $\ll 100$ mbar, snow is most stable in topographic lows (Fastook et al.,
 439 2008). When $P_o \gg 100$ mbar, snow is most stable on mountaintops.

440 Melting and runoff depend on energy fluxes around the hottest part of the day.
 441 Figure 9 shows the terms in the energy balance for a snow surface artificially
 442 initialized just below the freezing point. At low P , L_{fr} exceeds insolation and
 443 melting cannot occur. At high P , L_{fr} is much less important. Instead, absorbed
 444 insolation and greenhouse warming are balanced principally by radiative losses
 445 and melting. Whether or not surface melting can be sustained will depend on
 446 the partitioning of the subsurface absorbed insolation (Figure 9).

447 Setting $\Delta T = 5\text{K}$ raises peak melt rate from $0.44 \text{ kg/m}^2/\text{hr}$ to $1 \text{ kg/m}^2/\text{hr}$.
 448 Melt fraction reaches 1 in the shallow subsurface. Melt is produced for 6 (in-
 449 stead of 4) Mars-hours per sol, and there is some melt in the subsurface for
 450 7.5 (instead of 4) Mars-hours per sol.

451 Both α and \mathbf{O}' affect the energy absorbed by the snowpack. This can be shown
 452 by considering the energy absorbed by equatorial snow at equinox:

$$E_{\text{equinox}} \approx (1 - \alpha) \underbrace{L_\tau \left(\frac{1 - e^2}{1 + e \cos \Psi} \right)}_{\text{distance from Sun}}^{-2} \quad (4)$$

453 where E_{equinox} is the sunlight absorbed at noon at equinox, L_τ is the solar
 454 luminosity at Mars' semi major axis at geological epoch τ , Ψ is the minimum
 455 angular separation between L_p and either $L_s = 0$ or $L_s = 180$ (Murray and
 456 Dermott (2000), their Equation 2.20), and the atmosphere is optically thin. If e
 457 is large then peak insolation need not occur at equinox. Equation 4 shows that
 458 moving from average orbital conditions ($e = 0.06$, $\Psi = 90$) to optimal orbital
 459 conditions ($e = 0.15$, $\Psi = 0$) has the same effect on E_{equinox} as darkening from
 460 albedo 0.28 to albedo zero.

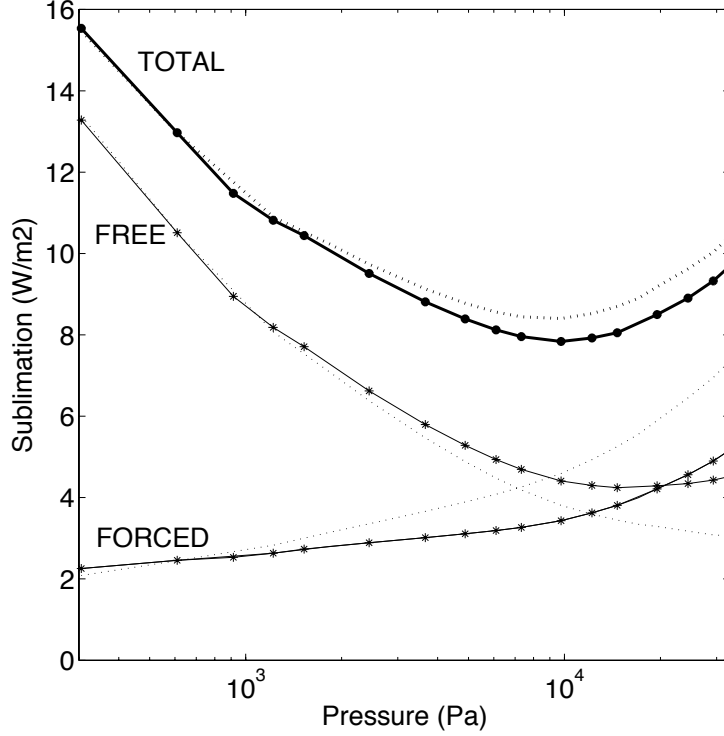


Fig. 8. Pressure dependence of sublimation rate at an equatorial site. FREE is L_{fr} , FORCED is L_{fo} . Solid lines with asterisks correspond to a wind speed that declines with increasing P , dotted lines correspond to constant near-surface wind speed of 3.37 m/s. $e=0.11$, $\alpha \approx 0.28$, $L_p=0$, $L_s=0$, $\phi=50$.

5.2 Seasonal cycle and snow locations

Figure 6 shows the seasonal cycle of T and sublimation rate on a flat planet. Annual average sublimation rate controls warm-season snow location, and annual-peak snow temperature determines whether melting will occur. The cold trap latitudes indicated correspond to $f_{snow} \rightarrow 0$, i.e. a single thick ice-sheet. Suppose instead that warm-season snow covers a wider area – that the “potential well” of Figure 7 fills up with snow. For modern orbital conditions, the area of snow stability will then extend south from the North Pole. If warm-season snow covers more than 43% of the planet – if the cold-trap effect is weak or nonexistent – then melt is possible even under modern orbital conditions. For optimal orbital conditions, increasing f_{snow} spreads the melting area to form a broad band equatorward of 30° .

5.3 Flat Mars – snowmelt on a cueball

Next, maps are summed over orbital states to find the latitudinal distribution of surface liquid water at a given P . If liquid water supply limits sedimentary

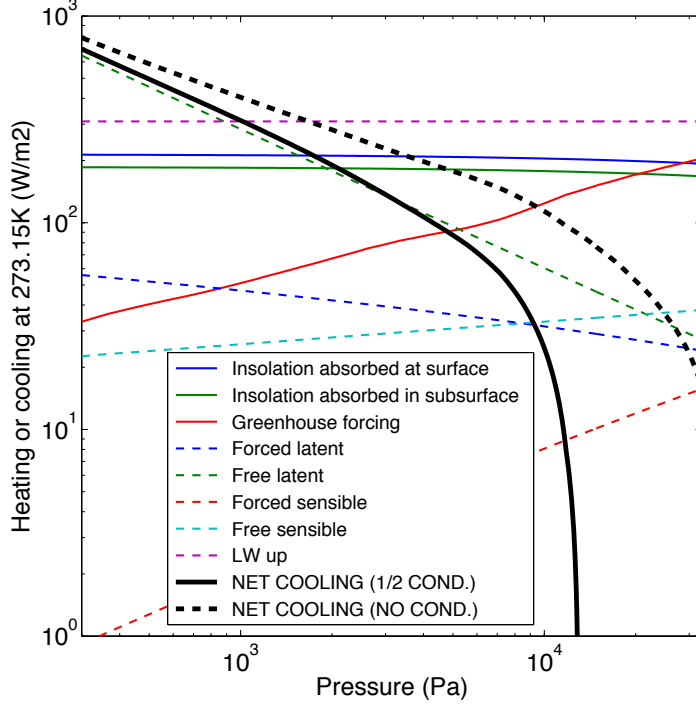


Fig. 9. Pressure dependence of the surface energy balance for an equatorial site with $T_1 = 273.15\text{K}$ imposed. Wind speed declines with increasing P . Note that subsurface melting can occur for $T_{surf} < 273.15\text{K}$. The absorbed insolation terms decrease slightly at high P due to Rayleigh scattering. Greenhouse forcing is stronger than in the time-dependent case shown in Figure 5 because the atmospheric temperature is assumed to be in equilibrium with a surface at the freezing point. If $1/2$ of the subsurface absorbed insolation returns to the surface through conductive heating, then melting will occur (net cooling at the melting point will be zero) for $P \gtrsim 130$ mbar. However, if none of the subsurface absorbed insolation conductively warms the surface, then surface melting will not occur even for the highest P shown (~ 330 mbar). S_{fo} and L_{fo} are weak because of the low surface roughness. $e=0.15$, $\alpha \approx 0.28$, $L_p=0^\circ$, $L_s=0^\circ$, $\phi=50^\circ$.

476 rock formation, then the sum over orbital states should correspond to geologic
477 observations of Early Mars deposits.

478 Suppose that the planet is flat, with warm-season snow only present in a
479 narrow ring at the latitude that minimizes the annually-averaged sublimation
480 rate. Suppose that we impose climate conditions only just allow melting under
481 the optimal orbital conditions. Then:–

482 *What is the latitudinal distribution of snow, melt and melt intensity?* Obliquity
483 is the strongest control on Mars snowpack stability. The 1D model predicts
484 that snow is most stable near the equator for $\phi \geq 40^\circ$, near the poles at $\phi =$
485 $\{0^\circ, 10^\circ, 20^\circ\}$, and at intermediate latitudes ($\pm 55^\circ$) for $\phi = 30^\circ$, in agreement
486 with other studies (Jakosky and Carr, 1985; Mischna et al., 2003; Levrard

et al., 2004; Forget et al., 2006; Madeleine et al., 2009). Nonzero e drives snow to the hemisphere in which aphelion occurs during summer. Holding e fixed, the width of the latitudinal belt swept out by warm-season snow during a precession cycle decreases with increasing ϕ , from $\pm 26^\circ$ at $\phi=40^\circ$ to $\pm 6^\circ$ at $\phi=80^\circ$ (for $e=0.09$). Holding ϕ fixed, the width of the warm-season snow belt increases with increasing e , from $\pm 18^\circ$ at $e=0.09$ to $\pm 22^\circ$ at $e=0.16$ (for $\phi = 40^\circ$).

99% of melting occurs for latitudes $<10^\circ$. Annual column snowmelt is sharply concentrated at the equator within the thin melt band (Figure 10). Even though the probability of a melt year is just $\sim 0.05\%$, the orbitally integrated expectation for the equatorial snowmelt column is 5 km/Gyr, which is the global spatial maximum on this flat planet. A typical Mars-year of melting produces 9 kg/m² melt at the latitude of warm-season snow. Peak instantaneous melt rate is of order 0.1 kg/m²/hr, and is less sharply concentrated at the equator than annual column snowmelt (Figure 10). Runoff is unlikely for these low melt rates, so any aqueous alteration must result from infiltration or from alteration of silicates within the snow. Low rates of infiltration may be sufficient to alter aeolian deposits, which volumetrically dominate the Mars sedimentary rock record (Grotzinger and Milliken, 2012).

What is the distribution of melt and melt intensity with orbital conditions? As **C** is moved towards conditions that allow melt, melt first occurs at $\phi \geq 40^\circ$ and $e > 0.13$ (Figure 11). Kite et al. (2011b,c) explain that high ϕ is needed to drive snow to the equator, and high e is needed to bring perihelion close to the Sun (Figure 11). Melting requires that perihelion occurs when the noontime sun is high in the sky – for the equator, this requires $Lp \sim 0^\circ$ or $Lp \sim 180^\circ$. Holding e and ϕ fixed and moving Mars through a precession cycle, the cueball planet is entirely dry between $Lp=45^\circ$ and $Lp=135^\circ$ inclusive, and between $Lp=225^\circ$ and $Lp=315^\circ$ inclusive.

What is the seasonal distribution of melt? All melting occurs near perihelion equinox. The melt season lasts ≤ 50 sols (for $e \leq 0.145$).

How does increasing f_{snow} affect the results? Pinning snow to $\pm 1^\circ$ of the optimum latitude corresponds to $f_{snow} \sim 1\%$, similar to the present-day value for warm-season snow. However, midlatitude pedestal crater heights (44 ± 22.5 m; Kadish et al. (2010)) are geologic evidence for high f_{snow} in the Late Amazonian. The modern surface ice reservoir is $2.9 \pm 0.3 \times 10^6$ km³ (Selvans et al., 2010; Plaut et al., 2007). Assuming secular loss of ice is slow, and that the midlatitude pedestal craters correspond to the thickness of a single ancient ice layer of uniform thickness, the ice accumulation area is $\sim 46\%$ of the planet's surface area. This ice accumulation area must be less than f_{snow} . At $f_{snow} = 50\%$ in the model, the melt belt thickens to $\pm 33^\circ$, with minor melt activity around $\pm 50^\circ$ (Figure 10). Maximum annual column melt and melt rates are

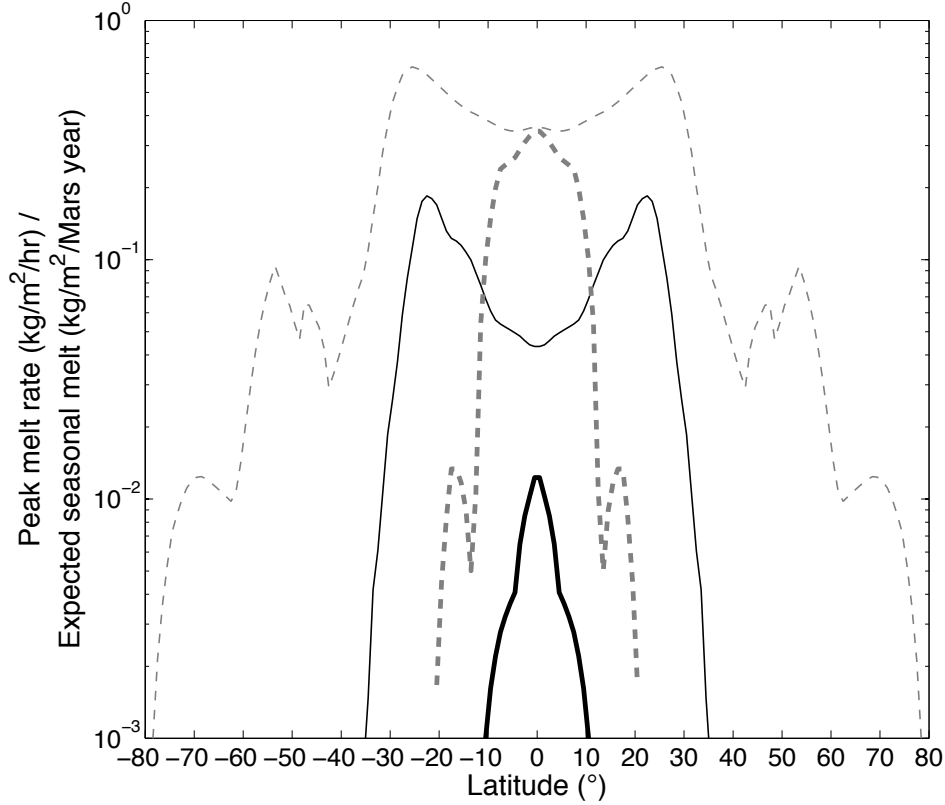


Fig. 10. Flat-planet results for a climate that only marginally rises above the freezing point on Early Mars. The black solid lines correspond to the orbitaly-averaged melt column per Mars year (relevant for aqueous alteration). The gray solid lines correspond to the peak melt rate experienced at any point in the orbits considered (the geomorphically relevant melt rate). The thick lines are for a very small value of f_{snow} (4%), and the thin lines are for $f_{snow} = 50\%$. The melt events poleward of 35° for $f_{snow} = 50\%$ correspond to extremely improbable orbital conditions.

at $\pm 22^\circ$, where column melt is $5\times$ greater (98 km/Gyr) and peak melt rates
are $3\times$ greater than at the equator. Peak melt rates now reach 0.7 mm/hr, so
runoff is conceivable (Figure 10). Although there is still a very strong increase
of melting with increasing e , melting can now occur for any value of L_p and
most e (given $\phi \geq 40^\circ$). Melt is strongest in the Northern Hemisphere at 15°
 $< L_p < 45^\circ$ and $135^\circ < L_p < 165^\circ$. It is strongest in the Southern Hemisphere
at $195^\circ < L_p < 225^\circ$ and $315^\circ < L_p < 345^\circ$. The explanation for this behaviour
is that the now-broad belt of warm-season snow never entirely shifts out of
the hot zone. The edge of this belt closest to the perihelion summer pole sees
the sun at zenith shortly before (and shortly after) perihelion solstice equinox.
This dramatic increase in melting does not require any change in greenhouse
forcing or paleopressure, just a change in the way the climate system deposits
snow.

How does pressure change affect snowmelt on a flat planet? At the highest P

considered, 293 mbar, low values of f_{snow} produce a broad band of melting between $\pm(15-20)^\circ$. There is a secondary peak around $\pm 50^\circ$. The maximum in melt rate snaps away from the equator at $f_{snow} \geq 35\%$, and above 35% this maximum moves to gradually higher latitudes. These patterns are similar at 98 mbar and 149 mbar, although the peak melt rates and expected-seasonal-mean melt rates are both lower because of the reduced greenhouse effect. At 49 mbar, melting only occurs for $f_{snow} > 35\%$, and never at equatorial latitudes. Lower P further restricts melting to high f_{snow} and higher latitudes. Physically, these trends correspond to the need for a long day-length (or perpetual sunlight) to warm the snowpack at low P . This is not possible at the equator where the day is always ≈ 12 hours long. The melt rate under orbital conditions that are optimal for melting can be thought of as a potential well in latitude, with maxima at high latitudes (for high- ϕ polar summer), and a minimum near the equator. At low P the melt potential is zero at low latitudes, so large values of f_{snow} are needed for melting, which will then occur away from the equator. Increasing P favors melting, so melt rates increase everywhere, and for 98 mbar and above the melt potential is nonzero even at the equator, allowing melting as f_{snow} nears 0. If Mars had a relatively thick $O(100 \text{ mbar})$ atmosphere during the sedimentary-rock era, then the progressive inhibition of equatorial melting is a latitudinal tracer of Mars atmospheric escape.

The “cueball” results presented in the above sections have strong echoes in the geologic record of ancient Mars. The distribution of evidence for liquid water shows strong latitudinal banding: sedimentary rocks are concentrated at latitudes $< 15^\circ$ (§3.1) but have “wings” at 25-30S and 20-30N, and alluvial fans are most common at 15S-30S (Kraal et al., 2008; Wilson et al., 2012).

5.4 With MOLA topography

The main meridional trends in model output without topography hold when MOLA topography is used. This is because nominal model parameters produce snow distributions that are more sensitive to latitude than to elevation.

At $P_o = 48 \text{ mbar}$ and optimal \mathbf{O}' ($\phi=50^\circ$, $e=0.16$, $L_p = 0^\circ$), snow distribution shows only a weak preference for low points. However, peak temperature is controlled by elevation because L_{fr} is stronger (and $LW \downarrow$ weaker) at high elevations. Therefore melting only occurs in planetary topographic lows close to the equator at low f_{snow} and $\Delta T=2\text{K}$ (Figure 12).

For $f_{snow} = 50\%$, melting at optimal orbital conditions occurs for all low-lying locations equatorward of 30° . When perihelion is aligned with solstice at $\phi=50^\circ$, the snow distribution shifts away from the equator, and no melting occurs below $f_{snow} \sim 60\%$.

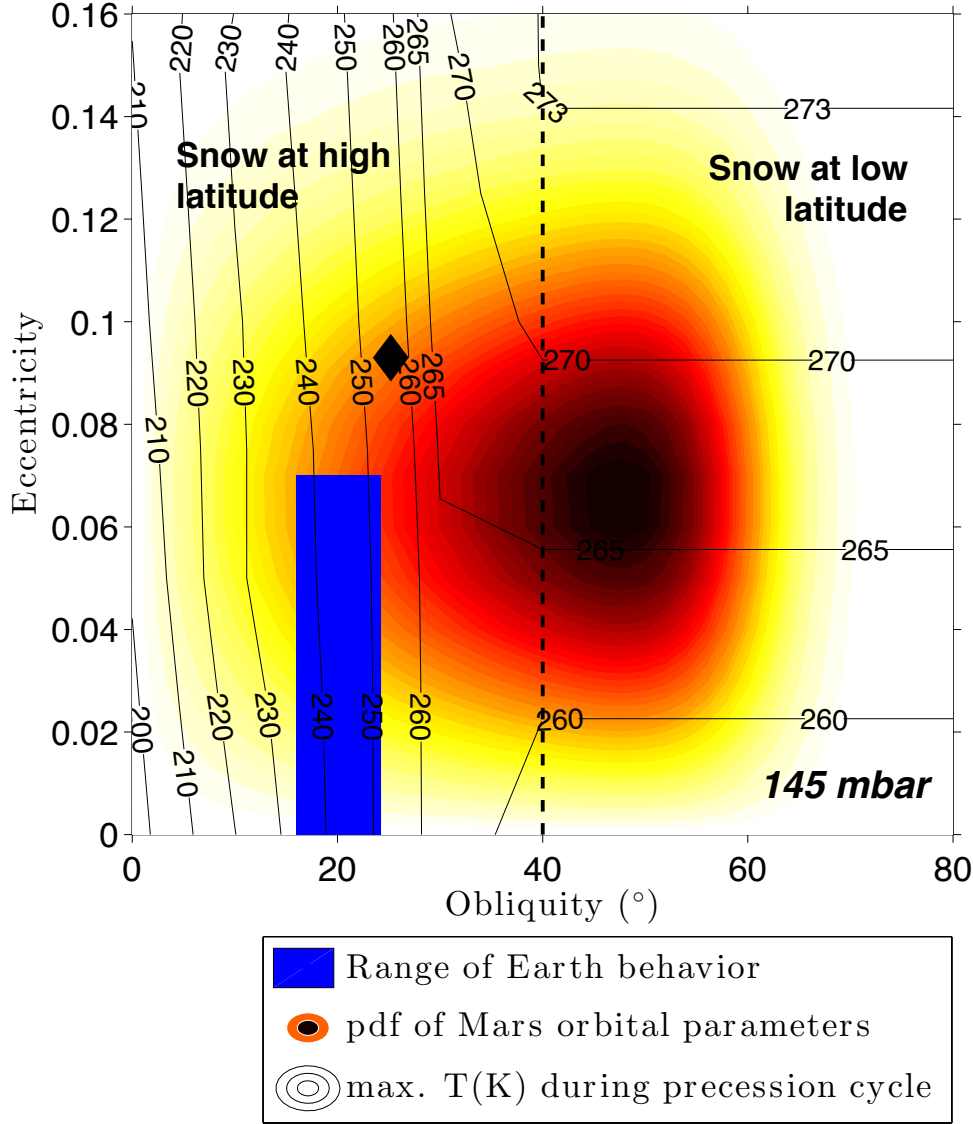


Fig. 11. The sensitivity of annual-peak snowpack temperature to Milankovitch forcing on an idealized Mars lacking topography. Maximum snowpack temperatures over a precession cycle (black contours) are highest for high obliquity and moderate eccentricity. The probable range of Mars orbital elements (color ramp, with white shading least probable and red shading most probable) is much broader than that of Earth's orbital elements (Gyr range shown by blue rectangle). Black diamond corresponds to Mars' present-day orbital elements. Vertical dashed line divides $\phi < 40^\circ$ (for which warm-season snow is generally found at high latitude), from $\phi \geq 40^\circ$ (for which warm-season snow is generally found at low latitude). $\Delta T = 5\text{K}$, $P = 145\text{ mbar}$, $\alpha \approx 0.28$, Faint Young Sun.

580 Low ϕ is much less favorable for snow melting than high ϕ (Jakosky and Carr,
581 1985). For $\phi \leq 30^\circ$ and perihelion aligned with solstice, snow is most stable
582 poleward of 60° , but these most favored locations never reach the freezing
583 point. As f_{snow} is raised, melting will first occur at lower latitudes because

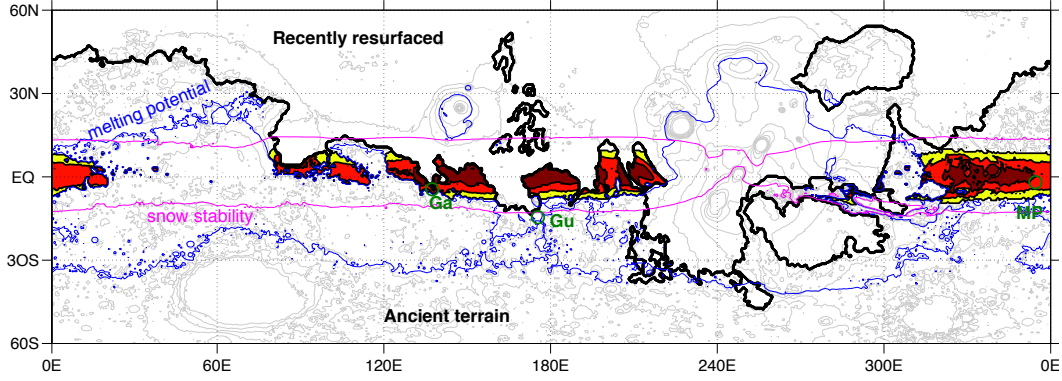


Fig. 12. A snapshot of snowmelt distribution for a single example of orbital forcing, showing role of snow stability and melting potential. $\phi=50^\circ$, $e=0.145$, $L_p = 0^\circ$, $P_o = 49$ mbar, $\Delta T = 5$ K. Areas equatorward of the magenta line correspond to $f_{\text{snow}} < 20\%$ – likely snowpack locations. Areas poleward of the blue line are hot enough to see melting at some point during the year, if snow were present. Where the snow zone intersects the hot zone, some melting will occur. Melt zones for small values of f_{snow} are shaded in warm colors. $f_{\text{snow}} < 10\%$ is shaded yellow, $f_{\text{snow}} < 5\%$ is shaded red, and $f_{\text{snow}} < 2\%$ is shaded maroon. Notice that extensive melting in Valles Marineris requires high f_{snow} or a different phase of the precession cycle. Thick black line corresponds to the boundary of terrain resurfaced since sedimentary rocks formed. This terrain is not included in the warm-shaded areas. Landing sites of long-range rovers are shown by green circles: – Ga = Gale Crater; Gu = Gusev Crater; MP = Meridiani Planum. Grayscale contours in background are topographic contours at intervals of 1.5 km from -5 km up to +10 km.

584 these receive more sunlight. The most favored locations are S Hellas and the
 585 lowest ground around 40N. These are the midlatitude locations where scal-
 586 loped depressions are most prominent (Soare et al., 2007; Zanetti et al., 2010),
 587 although these features might not require liquid water to form (Lefort et al.,
 588 2009, 2010) and the model is not directly applicable to Upper Amazonian
 589 features.

590 Snow and melt distributions on MOLA topography depend on the trade-off
 591 between P and sublimation rate, which controls snow stability (Figure 8). For
 592 example, suppose wind speed on Early Mars was much higher than modelled.
 593 Then the relative importance of wind-speed-dependent turbulent losses in the
 594 surface energy balance would increase. This would increase the importance of
 595 elevation ($\sim 1/P$) in setting snow location, relative to latitude which sets $SW\downarrow$.
 596 The snow and melt distributions for this “windy early Mars” (not shown) are
 597 broader in latitude and more concentrated in low areas (especially Northern
 598 Hellas, but also Northern Argyre and the Uzboi-Ladon-Margaritifer corridor).

600 Summing the orbital-snapshot maps of melt likelihood ($\int p(\mathbf{O}) d\mathbf{O}$) shows the
 601 effect of $\mathbf{C} = \{P, \Delta T, f_{\text{snow}}\}$ on melt likelihood averaged over geological time.

602 For $P_o = 49$ mbar and for small values of ΔT (5K) and f_{snow} (2%), warm-
 603 season snow is found primarily in (Figure 13a) Valles Marineris, the circum-
 604 Chryse chaos, the Uzboi-Ladon-Margaritifer corridor, craters in W Arabia
 605 Terra, the Isidis rim, northern Hellas, Gale, Aeolis-Zephyria Planum, and parts
 606 of the Medusae Fossae Formation, as well as at high ($>50^\circ$) latitudes. However,
 607 warm-season snow only melts very close to the equator (Figure 13b) – in
 608 Gale, the circum-Chryse chaos, Meridiani Planum, Aeolis-Zephyria Planum,
 609 the Isidis rim, and the floors of the Valles Marineris canyons. Even in central
 610 Valles Marineris, among the wettest parts of the planet under this climate,
 611 melting occurs with probability $<0.5\%$ (e.g., 5 Myr of melt years during 1
 612 Gyr). As f_{snow} is increased to 5-10% at $\Delta T = 5\text{K}$, melting in Meridiani Planum
 613 and Valles Marineris becomes more frequent. Melting in Northern Hellas does
 614 not occur until either f_{snow} or ΔT is greatly increased.

615 As the atmosphere is lost, melting becomes restricted in space as well as
 616 time ($\Delta T = 5\text{K}$, $P_o = 24$ mbar, $f_{\text{snow}} = 0.1\%$, Figure 14a). The last hold-
 617 outs for surface liquid water on Mars are Gale Crater, du Martheray Crater,
 618 and Nicholson Crater in the west-of-Tharsis hemisphere, and the floors of the
 619 Valles Marineris canyons in the east-of-Tharsis hemisphere (Figure 14a). Gale
 620 Crater (near 6S, 135E) is usually a hemispheric maximum in snowmelt for
 621 marginal-melting climates. Melting can only occur for very improbable orbital
 622 combinations under this climate. If they occurred at all, wet periods would be
 623 separated by long dry intervals.

624 At $P_o = 293$ mbar and low f_{snow} , low-latitude snow is restricted to high ground
 625 and so is melt. Figure 14b shows the melt distribution for $f_{\text{snow}} = 10\%$ and
 626 $\Delta T = 7.5\text{K}$. For $f_{\text{snow}} \geq 20\%$, snow is still most likely at high ground, but the
 627 melt pattern flips: melt occurs at all elevations, but it is most common at low
 628 ground as in the low- P case. As $f_{\text{snow}} \rightarrow 100\%$, melt extent is limited only
 629 by temperature. This is maximized at low elevations, because of the greater
 630 column thickness of greenhouse gas (and because of the adiabatic atmospheric
 631 temperature lapse rate, neglected here).

632 Hellas is usually the most favored area for snowmelt within the midlatitude
 633 ancient terrain. A climate relatively favorable for melting in Northern Hellas
 634 is shown in Figure 14c ($\Delta T = 15\text{K}$, $P_o = 24$ mbar, $f_{\text{snow}} = 40\%$). The contour
 635 of locally maximal snowmelt extending from deepest Hellas, to the Northern
 636 Hellas floor, to crater Terby, is intriguing in view of recent descriptions of thick
 637 packages of sedimentary rock in these locations (Wilson et al., 2007; Ansan

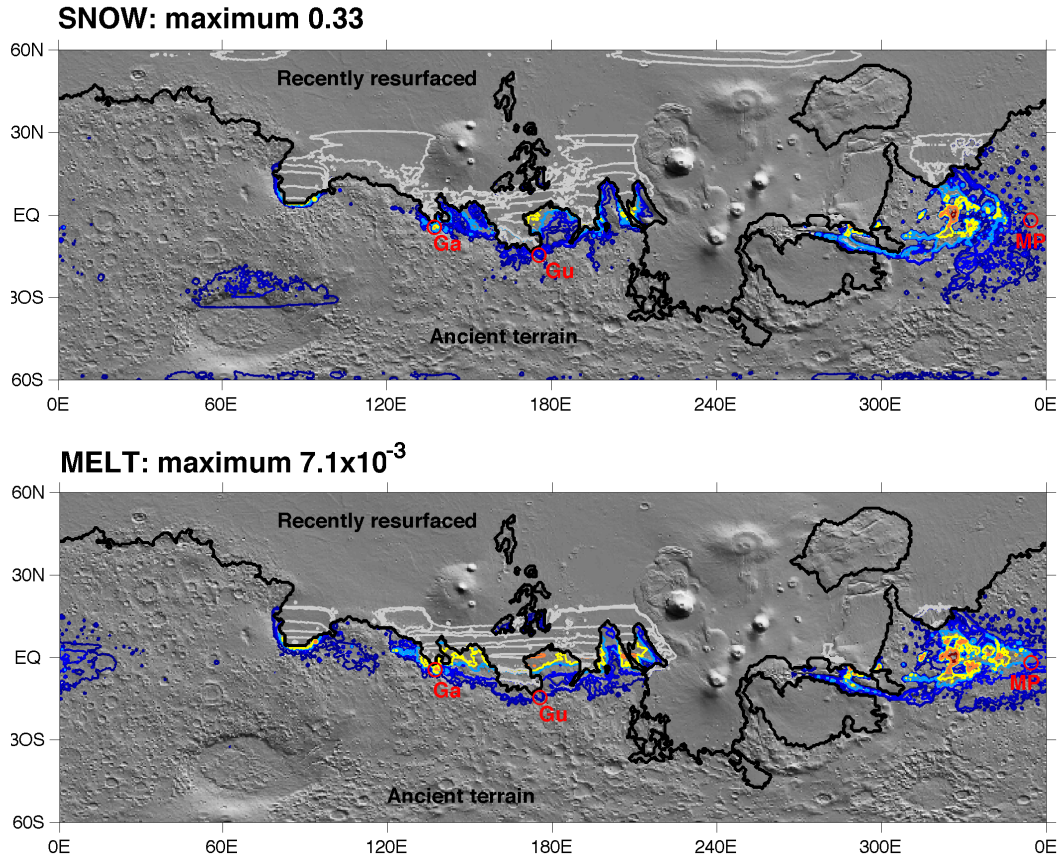
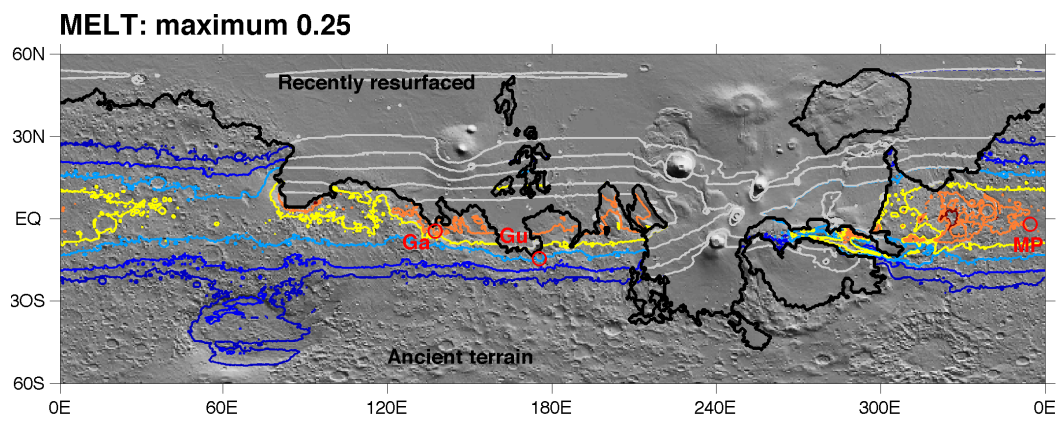
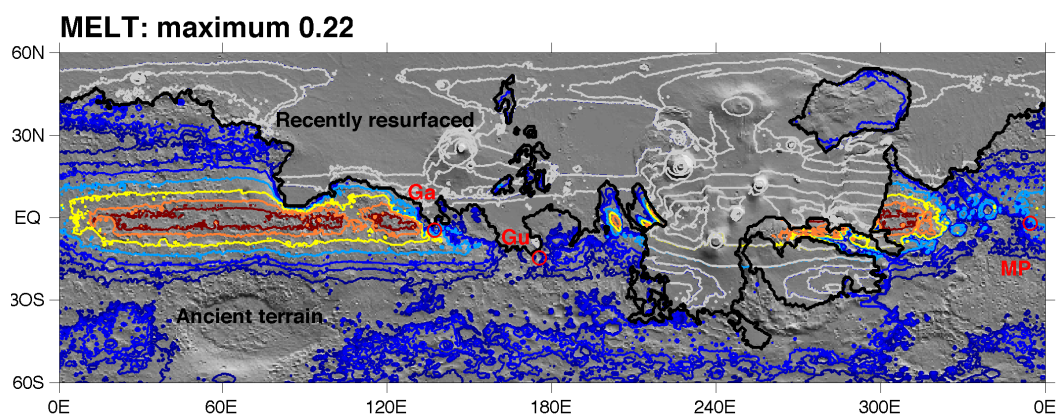
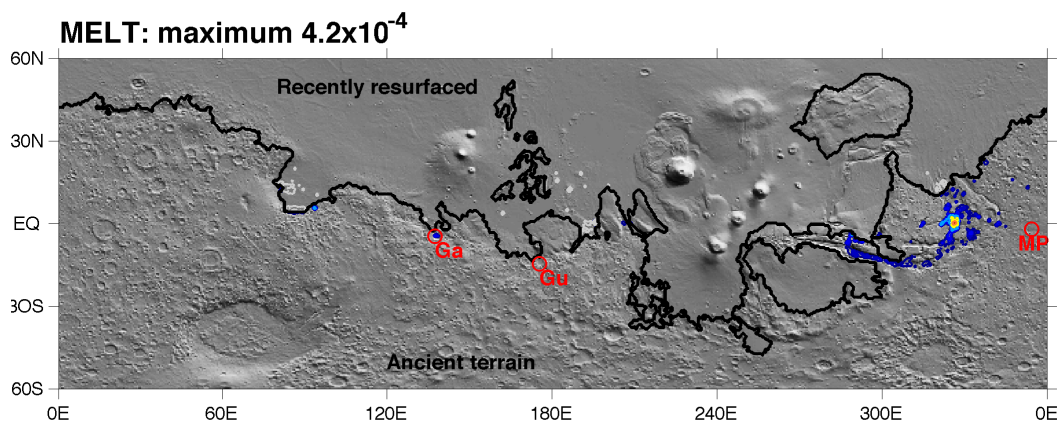


Fig. 13. Probabilities of (upper panel) warm-season snow and (lower panel) melting for $P = 49$ mbar, $\Delta T = 5\text{K}$, and $f_{\text{snow}} = 2\%$. Background is shaded relief MOLA topography, illuminated from top left. Maximum probability on the warm-season snow map is 0.33, maximum of the melt map is 7.1×10^{-3} – the location is Hydaspi Chaos for both maxima. Contours are at 1%, 5%, 10%, 25%, 50%, 75% and 90% of the maximum value. Because melting requires unusual orbital conditions, while warm-season low-latitude snow only requires high obliquity, the lowest colored contour in the snow map is greater than the highest colored contour in the melt map. Thick black line corresponds to border of ancient terrain, and grayed-out contours are probabilities on recently resurfaced terrain. Long-range rover landing sites are shown by red circles:– Ga = Gale Crater; Gu = Gusev Crater; MP = Meridiani Planum.

et al., 2011; Wilson et al., 2010). The Uzboi-Ladon-Margaritifer corridor of fluvial activity (Grant and Parker, 2002; Grant et al., 2008; Milliken and Bish, 2010; Mangold et al., 2012) is not as favorable for snowmelt.

For the wettest conditions considered (e.g., $\Delta T = 15\text{K}$, $P_o = 49$ mbar, $f_{\text{snow}} = 40\%$, Figure 14d), melt occurs more than 25% of the time in most places equatorward of 30° . Such wet global climates grossly overpredict both the spatial extent of sedimentary rock formation on Early Mars and the extent of surface aqueous alteration, as discussed in the next section.



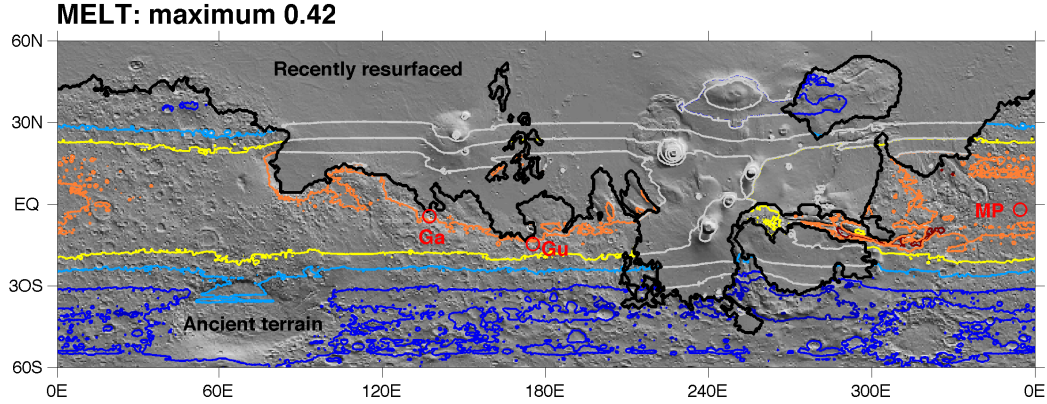


Fig. 14. Sensitivity of snowmelt maps to extreme variations in model parameters. Background is shaded relief MOLA topography, illuminated from top left. Colored contours correspond to snowmelt probabilities on ancient terrain. Contours are at 1%, 5%, 10%, 25%, 50%, 75% and 90% of the maximum melt likelihood, which is given to the top left of each panel. Black line corresponds to border of ancient terrain, and grayed-out contours are snowmelt probabilities on recently-resurfaced terrain. Long-range rover landing sites are shown by red circles:– Ga = Gale Crater; Gu = Gusev Crater; MP = Meridiani Planum. **(a)** Parameters that only marginally allow melting even under optimal orbital conditions: $P = 24$ mbar, $\Delta T = 5$ K, $f_{snow} = 0.1\%$. **(b)** High P drives snow (and melt) to high ground: $P = 293$ mbar, $\Delta T = 7.5$ K, $f_{snow} = 10\%$. This is inconsistent with the observed concentration of sedimentary rock at low elevations, but may be relevant to the distribution of older valley networks. **(c)** Parameters that produce snowmelt in Hellas: $P = 24$ mbar, $\Delta T = 15$ K, $f_{snow} = 20\%$. For legibility, the 1% contour is not shown for this subfigure. **(d)** Very high f_{snow} and ΔT predict a latitudinally broader distribution of sedimentary rocks than observed: $P = 49$ mbar, $\Delta T = 15$ K, $f_{snow} = 40\%$.

646 6 Snowmelt in space: understanding the distribution of sedimen- 647 tary rocks on Mars

648 6.1 Comparison of global data to model output: implications for Early Mars 649 climate state

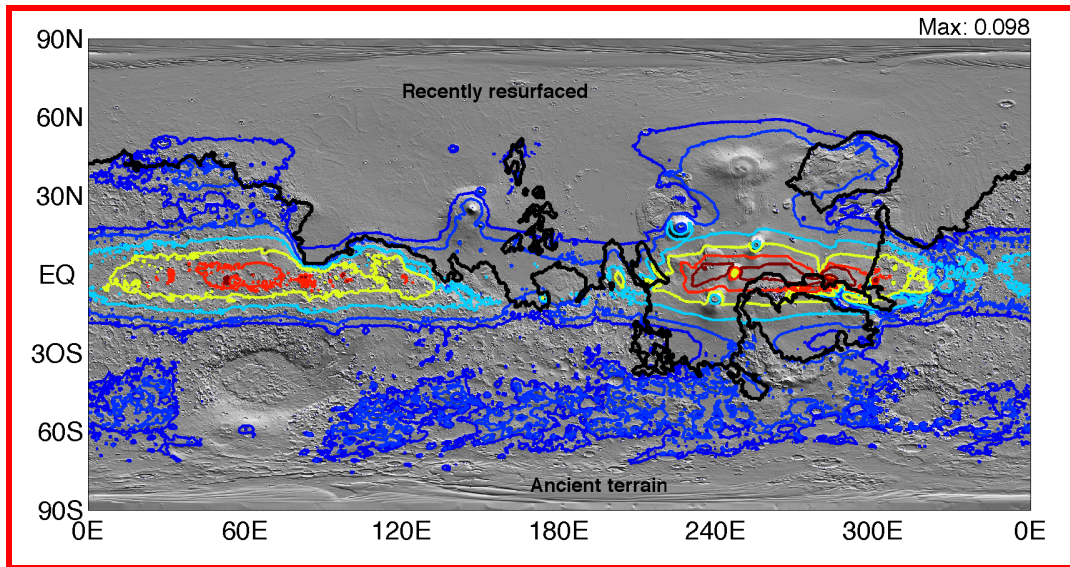
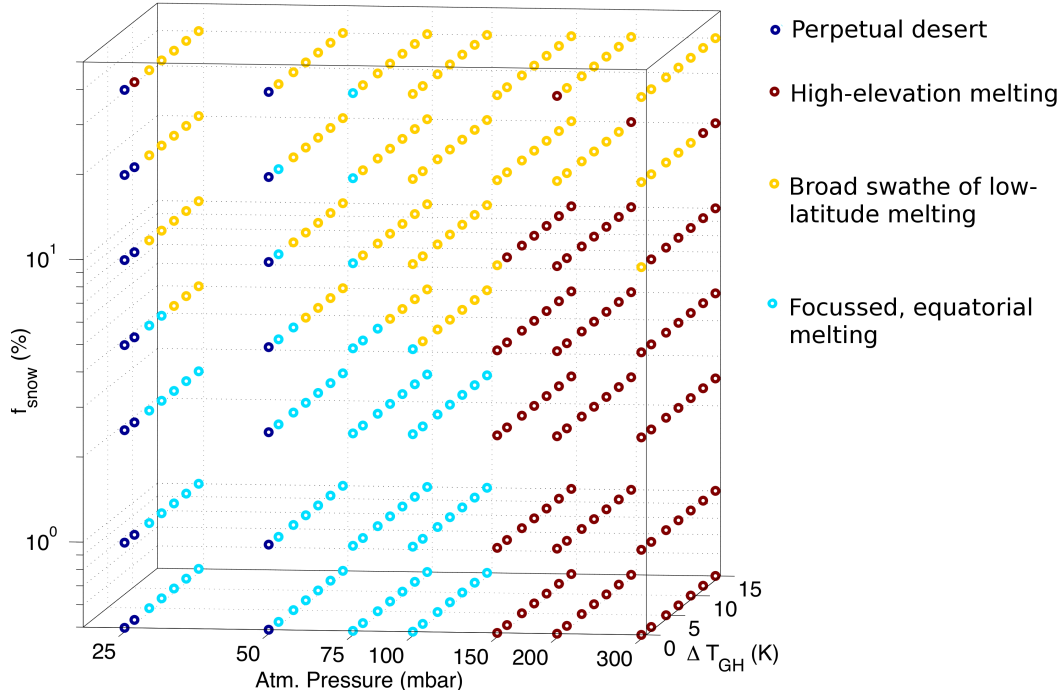
650 The full climate ensemble consists of 343 orbitally integrated melt-likelihood
651 maps similar to those in Figure 14. To reduce this to a manageable number
652 for analysis, k -means clustering was used (Press et al., 2007). The spatial
653 variability of the melt-likelihood maps was normalized by the within-map
654 mean and within-map standard deviation, and clustering was carried out on
655 these self-standardized maps. Representative results are shown in Figure 15,
656 together with the mean melt-likelihood maps for each of the climate clusters
657 identified.

658 Dark blue dots correspond to perpetual-global-desert climates. Zero melting
659 is predicted on horizontal surfaces under all orbital conditions.

660 At $P_o \geq 150$ mbar, all ΔT , and low-to-moderate f_{snow} , melting occurs at
661 high elevation (red cluster). Some melt also occurs at mid-southern latitudes.
662 Neglect of the adiabatic lapse rate will lead to growing inaccuracy at high P_o ,
663 but will not alter the conclusion that warm-season snow will be driven to high
664 ground at high P_o , far from the places where sedimentary rocks are observed
665 (Figure 15b).

666 For a wide range of P_o , all ΔT , and moderate-to-high f_{snow} , the framework
667 predicts a broad swath of low latitude melting (amber cluster). Figure 15c is
668 effectively a map of maximum snowpack temperature – as f_{snow} becomes large,
669 warm-season snow is no longer restricted by elevation. Melting is most intense
670 at low elevation because of the increased CO_2 column, but the overall pattern
671 is diffuse in both elevation and latitude. This contrasts with the strongly-
672 focussed observed sedimentary rock distribution (Figure 2). Melt probabilities
673 are large over a broad part of the planet, which does not sit easily with thermal
674 infrared data indicating that most soil on Mars did not experience volumetri-
675 cally important aqueous alteration (Bandfield et al., 2011).

676 For $P_o < 150$ mbar and at least one of low ΔT or low f_{snow} , the model pre-
677 dictes focused, equatorial melting (cyan cluster), in excellent agreement with
678 observations (Figure 15d). The agreement is especially good given the sim-
679 plicity of the model physics (§4) and the fact that we are considering one of
680 three objectively-defined *classes* of paleoclimates rather than the optimum **C**.
681 Independently of the snowmelt model, this result suggests that latitude and
682 elevation are the main controls on sedimentary rock distribution on Mars, be-
683 cause our model physics does not include 3D effects. We highlight seven points



of data/model agreement:–

- (1) The thickest sedimentary rock exposures on Mars are in Valles Marineris (up to 8km), Gale Crater (5km), and Terby Crater (3km) (Murchie et al., 2009a; Anderson and Bell, 2010; Wilson et al., 2007; Ansan et al., 2011). Sedimentary layered deposit thicknesses in the chaos source regions are up to ~ 1 km (Aram; Glotch and Christensen (2005)). The Medusae Fossae Formation is a sedimentary accumulation up to 3km thick (Bradley et al., 2002) which may also be aqueously cemented sedimentary rock in its lower part (Burr et al., 2009, 2010). With the exception of Terby,

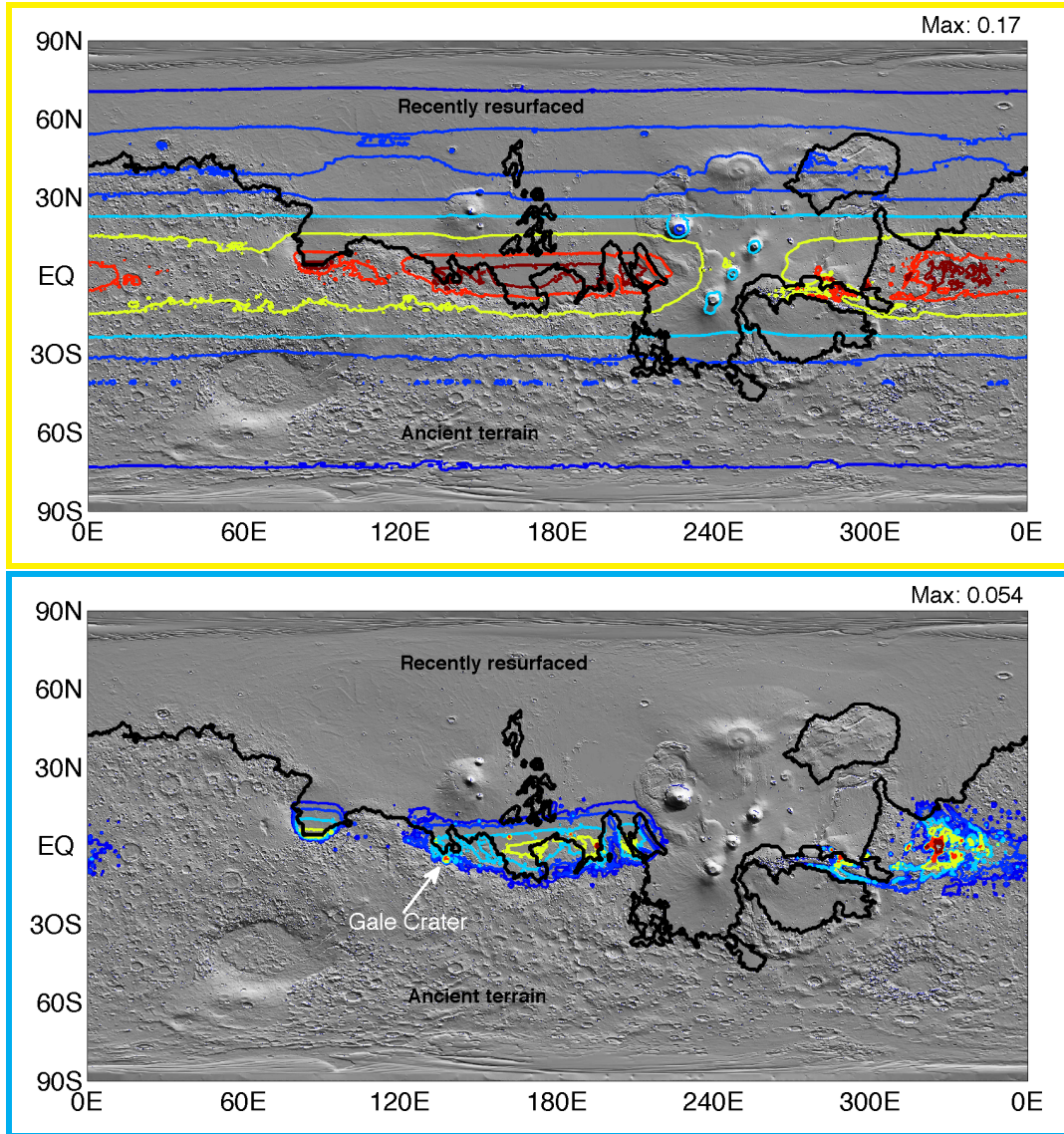


Fig. 15. Effect of climate on Mars sedimentary-rock distribution, assuming a snowmelt water source. Top panel shows clustering of climates into three melt-producing classes, plus perpetual-global-desert climates (dark blue dots). Remaining panels show maps of the mean of each of the melt-producing climate classes. Border colors correspond to the dots in the climate parameter space which contribute to that map. “Max” to the top right of each map refers to the spatial maximum in melt likelihood, which is the probability that a given location sees some melting during the year. The colored contours correspond to melt likelihoods of 5% , 10%, 25%, 50%, 75%, and 90% of the spatial maximum for that climate class. Thick black line shows the boundary between ancient terrain and recently-resurfaced terrain.

693 this is the same set of locations where the focussed, equatorial melt-
 694 ing paleoclimate class predicts maxima in orbitally-integrated snowmelt.
 695 The Northern Valles Marineris canyons contain thicker sedimentary-rock
 696 mounds than the southern Valles Marineris canyons, and are correspond-

- 697 ingly more favored for snowmelt in the model.
- 698 (2) Gale is a hemispheric maximum in ancient-terrain sedimentary rock thick-
699 ness, and is a hemispheric maximum in ancient-terrain snowmelt in the
700 model.
- 701 (3) Snowmelt is strongly focussed in the Valles Marineris, the chaos source
702 regions, and Gale. Predicted deposit thickness dies away quickly from
703 these regions.
- 704 (4) Meridiani Planum is correctly predicted to be a local maximum within a
705 broader wedge-shaped Sinus Meridiani outcrop narrowing and thinning
706 to the East (Edgett, 2005; Hynek and Phillips, 2008; Andrews-Hanna
707 et al., 2010; Zabusky et al., 2012; Wiseman et al., 2011). The concen-
708 tration of Western Arabia sedimentary rock in mound-filled craters (e.g.
709 Crommelin, Firsoff, Danielson, Trouvelot, and Becquerel) is reproduced
710 by the model. Alignment of Meridiani Planum with snowmelt maximum
711 implies net True Polar Wander $<10^\circ$ since sediment deposition (Mat-
712 suyama et al., 2006; Perron et al., 2007; Kite et al., 2009; Matsuyama
713 and Manga, 2010).
- 714 (5) The southern Isidis rim is identified as a regional maximum for post-
715 Noachian surface liquid water, consistent with data (eg., Jaumann et al.
716 (2010)).
- 717 (6) In the Northern Plains, deep equatorial craters are commonly modified by
718 sedimentary infill (e.g., Nicholson, Reuhl). This correlation is reproduced
719 by the model.
- 720 (7) The focussed equatorial melting climate cluster predicts strong enhance-
721 ment of melting in Northern Hellas relative to other locations in the
722 same latitude band (similar to Figure 14). However, $\Delta T \geq 10\text{K}$ is needed
723 for non-negligible melting away from the equator, so this longitudinal
724 enhancement is diluted in the class-average map and is not visible. A
725 secondary enhancement within this southern latitude belt is the Uzboi-
726 Ladon-Margaritifer corridor. These longitudinal enhancements match data
727 on the distribution of sedimentary rocks and alluvial fans (Kraal et al.,
728 2008). However, the model underpredicts the thickness of Terby fill, rel-
729 ative to the equatorial belt of sedimentary rocks.

730 The model predicts snowmelt in the circum-Chryse region out of proportion to
731 the observed sedimentary rock deposits. However, chaos and outflow channels
732 continued to form around Chryse through the Early Amazonian (Warner et al.,
733 2009; Carr and Head, 2010), and would have destroyed sedimentary rocks
734 deposited earlier. If supraglacial snowmelt crevassed to the base of the ice
735 mass and inflated subglacial lakes, seasonal melting could have contributed
736 to chasm flooding and overflow. Sedimentary rocks overly chaos in Aram and
737 Iani (Glotch and Christensen, 2005; Warner et al., 2011). The model does
738 not predict snowmelt at Mawrth, consistent with Mawrth's interpretation as
739 a Noachian deposit formed under a earlier climate (McKeown et al., 2009),
740 nor does it predict snowmelt at Terra Sirenum, consistent with the nominally

741 Late Noachian age of the inferred paleolake deposits there (Wray et al., 2011).
 742 Finally, the model predicts snowmelt mounds within deep Northern Lowlands
 743 craters that appear little-modified in CTX, such as the crater at 94.5E, 10N.
 744 If the snowmelt hypothesis is correct, then these craters must postdate the
 745 sedimentary-rock era in order to avoid infilling by sedimentary rock. This
 746 prediction can be tested with crater counts on ejecta blankets.

747 The colors assigned to the climate classes correspond to a hot-to-cold sequence
 748 in Early Mars climate parameter space. The high elevation (red) and broad-
 749 swath (amber) classes have melt likelihoods as high as 0.17. The focussed
 750 equatorial (cyan) class shows much lower melting probabilities (≤ 0.054) and
 751 is wrapped around the perpetual-global-desert climates (dark blue). To obtain
 752 the distribution of snowmelt that is in best agreement with sedimentary rock
 753 data, at least one of P_o , ΔT or f_{snow} must be small. This assumes that sedimen-
 754 tary rock accumulation is proportional to the number of years with snowmelt.
 755 Sedimentary rock formation involves nonlinear and rectifying processes, which
 756 could allow the snowmelt production predicted by the broad-swath-of-low-
 757 latitude melting (amber) class to yield a sedimentary rock distribution con-
 758 sistent with observations. Alternatively, a broad initial sedimentary-rock dis-
 759 tribution could be focussed by aeolian erosion, which would preferentially ef-
 760 face thin deposits. Therefore, data-model comparison supports the focussed-
 761 equatorial-melting climate class (cyan), and rules out the high-elevation melt-
 762 ing paleoclimates (red class) and the perpetual global desert (dark blue). It
 763 disfavors, but does not rule out, the broad-swath-of-melting (amber) climate
 764 class.

765 In summary, if the seasonal-melting hypothesis is correct, Mars paleoclimate
 766 has left a fingerprint in the sedimentary rock distribution. Sedimentary rocks
 767 are distributed as expected if Mars only marginally permitted snowmelt, even
 768 under near-optimal orbital conditions. The climates that give the best fit to
 769 data predict planets on which the wettest geographic location would have been
 770 dry for $\gtrsim 90\%$ of the time. During the sedimentary-rock era, Mars was a dry
 771 place.

772 6.2 Possible implications for other geologic data: valley networks, chlorides, 773 and alluvial fans

774 Regionally-integrated valley networks record overland flow *prior* to the sed-
 775 imentary rock era. We find that Mars valley-network elevation distribution
 776 is biased high by 600m relative to ancient terrain, although this may reflect
 777 the generally higher elevation of mid-Noachian (as opposed to Early Hesper-
 778 ian) outcrop (Hynek et al., 2010). High elevation is the fingerprint of high
 779 P_o (Figure 14b; Figure 15b). This suggests a geologic record of progressive

atmospheric loss:– P_o is >100 mbar at valley-network time (to drive snow to high ground as suggested by valley network elevations), falls to ~ 100 mbar by sedimentary-rock time (high enough to suppress evaporative cooling, low enough to allow sedimentary rock formation at low elevation), and falls further to the current situation (6 mbar: L_{fr} prevents runoff on horizontal surfaces). The melt rates predicted by our model with nominal parameters are \lesssim mm/hr, probably insufficient to form the classical valley networks. Processes that could link runoff from snowmelt to the formation of classical highland valley networks include:– (1) a stronger greenhouse effect than considered here, with or without the orbital variability considered in this paper; (2) increasing e to ~ 0.22 , as can occur transiently during the restructuring of solar system orbital architecture predicted by the Nice model (Agnor and Lin, 2012); (3) transient darkenings from impact ejecta and ash and transient heating from impact ejecta.

Chloride deposits ($n = 634$) are generally older than the sedimentary rocks, extremely soluble, rare in the equatorial sedimentary rock bracelet, and regionally anticorrelated with sedimentary rock (Osterloo et al., 2010). This excludes an erosional mechanism for the latitudinal distribution of sedimentary rocks. One possibility is that dust obscures chlorides at low latitudes. Another possibility is that chlorides were dissolved in the equatorial band during the melt events that lithified the sedimentary rocks. This would imply that melt rarely occurred far from the equator.

Peak runoff production *during* the sedimentary-rock era is constrained to $\sim 0.3 \pm 0.2$ mm/hr (Irwin et al., 2005; Jaumann et al., 2010). Melt production at these rates is possible in the climate-ensemble shown. However, runoff production will be some fraction of the melt rate, because of refreezing and infiltration. Similar to the case of the classical highland valley networks, additional energy could be supplied by transient darkenings (ash-on-snow, ejecta-on-snow) or transient volcanogenic warming. An alternative way to maximize runoff at low P_o is a phase lag between and the position of cold traps (which is set by orbital forcing, e.g. Montmessin et al. (2007)) and the position of ice deposits. For example, an ice deposit built up at 20S at high ϕ while $L_p \sim 90^\circ$ may melt if it is not removed by sublimation before L_p swings back to 270° . This phase lag contrasts to the snow considered in this paper, which is always in equilibrium with orbital forcing.

Evidence that the alluvial fans are *younger* than most sedimentary rocks (Grant and Wilson, 2011) is consistent with loss of CO_2 over time, because low P_o suppresses equatorial melting (§5.3). Figure 10 shows that at high f_{snow} on a cueball planet, melt rates peak at $\pm 22^\circ$, being negligible at higher latitudes and several-fold lower at the equator. The latitude of the wing “peak” increases with f_{snow} , but wings exist for a broad range of moderate-to-high f_{snow} and P_o . Wings are observed in the latitudinal distribution of alluvial

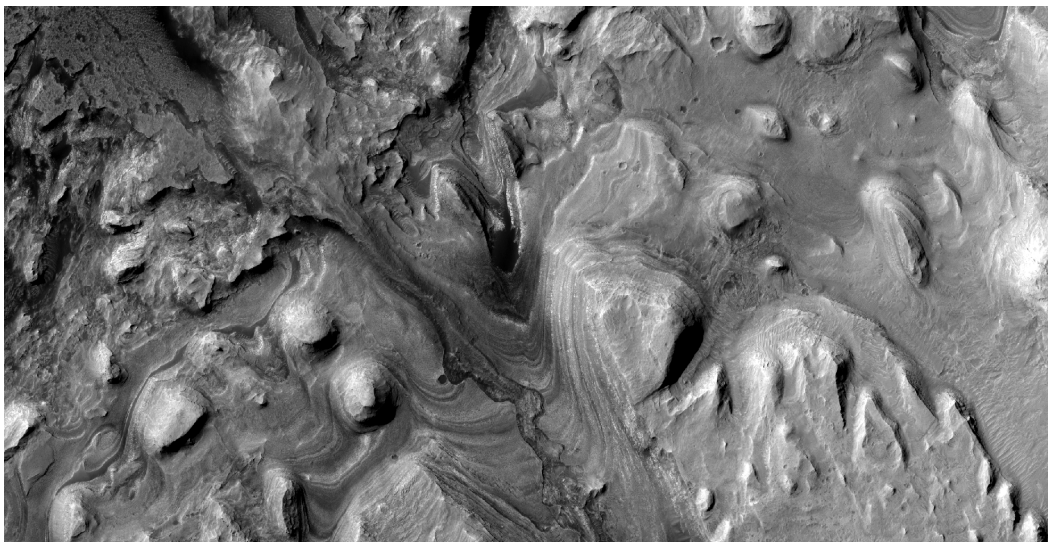


Fig. 16. Possible Mars Science Laboratory primary mission exploration targets in the foothills of the Gale Crater mound. Finely layered sulfate-bearing and phyllosilicate-bearing sedimentary rocks, locally reworked by a channel. See Anderson and Bell (2010) for geologic context. Part of HiRISE PSP_009294_1750. Image is ~4500 m across, illumination is from top left.

822 fans on Mars (Kraal et al., 2008; Wilson et al., 2012).

823 **7 Snowmelt in time: predictions for MSL at Gale Crater**

824 *7.1 Testing snowmelt at Gale Crater*

825 The base of the Gale mound is a good place to test the snowmelt hypothe-
 826 sis, because snowmelt is predicted at Gale for most of the paleoclimates that
 827 permit surface liquid water anywhere on Mars.¹ Model predictions are for
 828 the lower unit of the Gale Crater mound, which is known to contain aqueous
 829 minerals (Milliken et al., 2010), not the spectroscopically bland upper unit
 830 (Thomson et al., 2011) for which much less snowmelt is predicted. We think
 831 that it is not a coincidence that Gale is a good place to test the snowmelt
 832 hypothesis. MSL was sent to Gale because it hosts one of the thickest sed-

¹ Of the subset of climate states considered that predict snowmelt anywhere on the planet, 66% predict snowmelt at the base of the Gale Crater mound. If we say that Gale Crater has a “robustness” of 66%, then >99% of ancient surface area scores lower for robustness. In addition, for 55% of climates modeled, the base of the Gale Crater mound is in the top 1% of the planet for melt likelihood. If we say that the base of the Gale Crater mound has a “maximality” of 55%, then >99.9% of ancient surface area scores lower for maximality.

imentary rock packages on Mars, with mineralogic and stratigraphic hints of climate change (Milliken et al., 2010). The snowmelt model predicts relatively abundant snowmelt at Gale, even in a changing Early Mars climate. If snowmelt is the limiting factor in sedimentary-rock production, then naturally Gale would be a place that would sustain sedimentary-rock formation for the widest range of climate conditions.

Hypothesis: We hypothesize that the Gale Crater mound is an accumulation of atmospherically-transported sediments pinned in place and subsequently reworked by seasonal-meltwater-limited processes. (This hypothesis is adumbrated in a unpublished M.S. thesis by Cadieux (2011), and in conference abstracts by Cadieux and Kah (2011) and by Niles and Michalski (2012).)

Tests: The snowmelt model predicts:–

- *Wet-dry cycles on orbital timescales*, with dry conditions most of the time.
- *Mound-scale geochemistry records a succession of closed systems, not a flow-through geochemical reactor.* If the fluids responsible for alteration were in contact with the atmosphere, as is true of all the ancient waters yet sampled by meteorites and rovers (Halevy et al., 2011), then $\bar{T} < 273\text{K}$ implies restriction of diagenesis to perched aquifers within meters of the surface (more beneath lakes). The Gale Crater mound is 5km high, so this predicts that the Gale Crater mound is a succession of tens-to-thousands of closed systems. If on the other hand the layers near the top of the mountain were precipitated from groundwater that had flowed from the bottom of the mountain, then the mountain is a flow-through geochemical reactor. Basal layers would then be vulnerable to alteration by subsequent upwelling fluids. If smectite layers are found between Mg-sulfate layers, this would place a tight upper limit on flow-through aqueous chemistry (Vaniman, 2011).
- *Clay/sulfate transitions correspond to a change in silicate input, not a change in global environmental chemistry.* At Gale and many other sites on Mars, sedimentary rocks transition upsection from irregular to rhythmic bedding (Grotzinger and Milliken, 2012). This suggests a change over time in the relative importance of transient darkenings from volcanism and impacts, versus orbital forcing. Early on, large explosive eruptions and large impacts were more frequent – so many melt events were assisted by regional-to-global albedo reduction. As volcanism and impacts declined, darkening events became less frequent, so eccentricity change (Figure 11) emerged as the key regulator of melt events. Therefore, we predict that the phyllosilicate layers in the base of the Gale Crater mound were altered in-situ, and are stratigraphically associated with impact ejecta (Barnhart and Nimmo, 2011) or volcanic ash layers.
- *Generally homogenous chemistry and mineralogy on ascending the mound.* With the exception of these events, the protolith is globally-averaged atmospherically-transported sediment, and most alteration is local. This leaves little scope

- 875 for unmixing of major-element chemistry.
- 876 • *No Gale-spanning lakes (except immediately after the Gale-forming impact).*
 - 877 Local perennial lakes are possible, as in the Antarctic Dry Valleys (Doran
 - 878 et al., 1998).
 - 879 • *Isotopic gradients.* Within a unit representing a single identifiable melt
 - 880 event, isotopic trends will depend on the water loss mechanism. If the wa-
 - 881 ter evaporated, earlier deposits will be isotopically lighter (in H and O
 - 882 isotopes) and later deposits heavier. This is due to the preferential evapo-
 - 883 ration of light water and will give an O isotope trend similar to that seen
 - 884 within ALH84001. If, on the other hand, the water froze rather than evap-
 - 885 orated, later deposits will be lighter or no time dependent trend in isotopic
 - 886 composition will be observed. By contrast, in a groundwater model, if the
 - 887 supply of groundwater is \sim constant during mineralization, then the isotopic
 - 888 composition of the evaporating fluid will be some steady-state value, which
 - 889 would depend on the isotopic composition of the upwelling fluid and the
 - 890 evaporation rate. Lesser variability is expected within a single deposition
 - 891 event.
 - 892 • *No organic carbon.* Slow, orbitally-paced sedimentation and oscillation be-
 - 893 tween reducing and oxidizing conditions would disfavor preservation of or-
 - 894 ganic carbon.

895 7.2 From snowmelt time series to the Gale Crater stratigraphic logs

896 *Seasonal cycles and runoff.* Early in the melt season, melt will percolate ver-

897 tically and refreeze (Marsh and Woo, 1984). Vertical infiltration of snowmelt

898 can indurate and cement aeolian dust and sand. Draining and channeliza-

899 tion of melt will lengthen the lifetime of subsurface melt, especially late

900 in the melt season. The impermeable ice table constructed by refreezing of

901 early-season melt favors late-season runoff. Runoff and ponding of snowmelt

902 in ice-covered lakes requires that water reaches channels before it refreezes.

903 Once water reaches channels, ice cover protects against further freezing. Be-

904 cause the daily average temperature is below freezing (in general this is not

905 a requirement for seasonal-melting models, but it is a feature of the model

906 output considered here), this requires that drainage times through firn are

907 <1 sol, in turn requiring high drainage density. Channel deposits with high

908 drainage density are sometimes seen within the sedimentary rocks of Mars

909 (e.g., HiRISE ESP_020602_1755 and PSP_007474_1745), and feed into much

910 larger (and much more frequently preserved) inverted channels. A possible ter-

911 restrial analog for these processes is the Coastal Thaw Zone of the Antarctic

912 Dry Valleys.

913 *Milankovitch cycles.* Snowmelt predictions are mapped onto sedimentology

914 and stratigraphy in Figure 17 (compare Figure 1). Wet-dry cycles with pe-

915 riod ~ 20 Kyr are inevitable unless $\Delta T \sim 15\text{K}$. Early in the wet phase of a
 916 wet-dry cycle, infiltration can provide water for diagenesis of layers that were
 917 deposited under dry conditions (Figure 17). As cementation reduces perme-
 918 ability, infiltration will decline and runoff will be increasingly favored. The
 919 primary control on temperature cycles is precession, with secondary control
 920 by ~ 100 Kyr eccentricity cycles. The “steady accumulation” column in Fig-
 921 ure 17 shows sedimentological predictions for the case where atmospherically-
 922 transported sediment is lithified by infiltration of snowmelt. The “wet-pass
 923 filter/disconformities” column shows the case where rock formation only oc-
 924 curs during wet intervals. This produces major disconformities. Quasi-periodic
 925 liquid water availability at Gale will not necessarily produce quasi-periodic
 926 sedimentology. On Earth, orbitally-paced climate signals are recorded with
 927 high fidelity by abyssal sediments, but are shredded by fluvial processes and
 928 so are barely detectable in fluviodeltaic sediments (Pälike et al., 2006; Jerol-
 929 mack and Paola, 2010).

930 Sequence stratigraphy on Earth divides stratigraphic packages into periods
 931 of relative sea-level rise and fall. Snowmelt model output suggests that the
 932 equivalent of sequence stratigraphy for Mars will involve mound accumulation
 933 during rare wet periods, and mound degradation (and apron accumulation?)
 934 during more common dry periods.

935 8 Discussion

936 8.1 Validity of model assumptions

937 The snowmelt model assumes that the *gross* accumulation rate of atmospher-
 938 ically transported sediment integrated over 4 Ga is enough to build up thick
 939 sedimentary rocks everywhere on Mars, but that the supply of liquid water
 940 is the limiting step for *net* accumulation. Alternative limiting factors include
 941 sediment availability and preservation/exposure.

942 Malin and Edgett (2000, 2001) and Edgett and Malin (2002) have suggested
 943 sedimentary rocks were once much more widespread. In this case, restricted
 944 exposure today would correspond to blanketing or erosion since sedimentary-
 945 rock time.

946 Sediment availability is unlikely to have limited sedimentary rock formation on
 947 Mars, because gross deposition rates for atmospherically-transported sediment
 948 on today’s Mars ($10^{1-2} \mu\text{m}/\text{yr}$; Arvidson et al. (1979); Geissler (2005); Geissler
 949 et al. (2010); Johnson et al. (2003); Kinch et al. (2007); Drube et al. (2010))
 950 are not much less than past accumulation rates of sedimentary rocks: 20-50

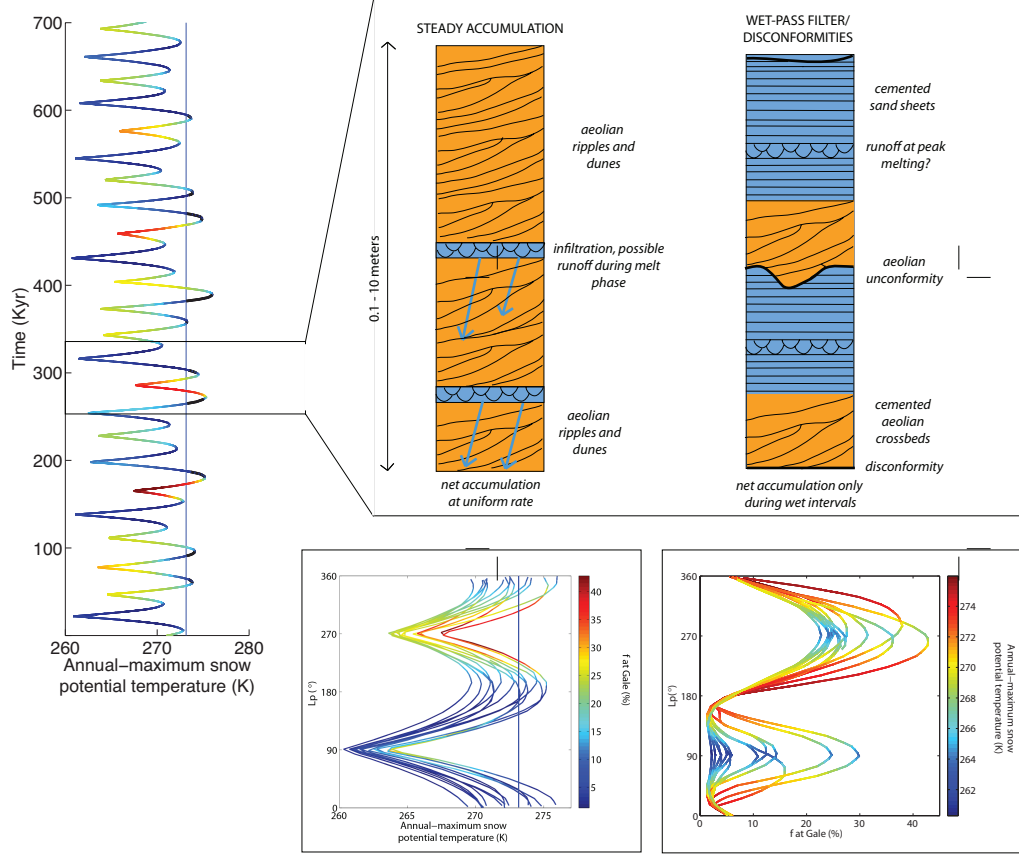


Fig. 17. Snowmelt model predictions for seven hundred thousand years at Gale Crater. *Left time series*: Potential temperature of snowpack at Gale. Because the Solar System cannot be deterministically reverse integrated to 3.5 Gya, the orbital forcing is necessarily fictitious, but it is realistic (Laskar solution 301003BIN_A.N006 for 73.05–73.75 Mya, but with 0.02 added to the eccentricity). The color scale corresponds to f at Gale. Red is unfavorable for warm-season snow at Gale, and blue is most favorable for warm-season snow at Gale. The vertical blue line corresponds to the melting threshold. Black highlights intervals of melting at Gale for $f_{\text{snow}} = 10\%$. $\Delta T = 6\text{K}$, $P_0 = 49$ mbar. *Stratigraphic logs*: Two end-member stratigraphic responses to orbitally-paced wet-dry cycles over ~ 50 Kyr interval. Orange corresponds to sediment accumulated during dry intervals, and blue corresponds to sediment accumulated during wet intervals. In the left column, the Gale Crater mound accumulates steadily with time, and layers are cemented by infiltration during wet intervals. In the right column, both accumulation and diagenesis are restricted to wet intervals. *Lower panels*: Precession cycles of temperature and f . Imperfect cyclicity results from varying eccentricity. Perihelion during northern-hemisphere summer is especially favorable for snow accumulation at Gale. Gale is dry when perihelion occurs during southern-hemisphere summer: snow accumulation is unlikely, and any snow that does accumulate fails to reach the melting point.

951 $\mu\text{m}/\text{yr}$ at the best-measured site (Lewis et al., 2008). Planet-wide sand motion
 952 occurs on Mars (Bridges et al., 2012). Sand transport and dust lifting on Mars
 953 is sensitive to small increases in P (Newman et al., 2005). Because Mars has

lost atmospheric CO_2 over time (Barabash et al., 2007), gross accumulation rates for atmospherically-transported sediment were probably $\geq 10^{1-2} \mu\text{m/yr}$ at the $O(10^2)$ mbar level required for snowmelt. Present-day reservoirs of air-fall sediment are large. For example, dust deposits at Tharsis and E Arabia Terra are $O(10)\text{m}$ thick (Bridges et al., 2010; Mangold et al., 2009). On Early Mars, background dust supply would be supplemented by sediment produced during impacts and volcanic eruptions. We do not have good constraints on the current surface dust (and sand) budget and how important finite dust reservoirs are for the current dust cycle, let alone on Early Mars. Therefore, applying current deposition rates to make the argument that sediment availability is not a limiting factor is fairly speculative. Nevertheless, if fluxes and reservoirs were as large in the past as today, then sedimentary rock formation would not have been limited by the availability of atmospherically-transported sediment. The difficulty then is to pin the sediment in place for >3.3 Gyr, and cementation by snowmelt is one mechanism that can resolve this difficulty.

Spatially varying precipitation is ignored. On Earth, “[a]dvances and retreats of glaciers are broadly synchronous” (Cuffey and Paterson, 2010), because small changes in Earth \mathbf{x} , \mathbf{O}' , and ΔT overwhelm regional variations in precipitation. This ablation sensitivity is what makes glaciers good dipsticks for Earth’s paleoclimate. On Mars, recent glaciations have laid down geomorphic strips near-parallel to lines of latitude, suggesting that longitudinally variable precipitation is less important than insolation in controlling precipitation and snowmelt (Kreslavsky and Head, 2000, 2003; Heldmann and Mellon, 2004; Neumann et al., 2003; Kadish et al., 2009; Hauber et al., 2008; Fassett et al., 2010). The most recent equatorial ice deposits formed at intermediate elevations on Tharsis and Terra Sabaea (Forget et al., 2006; Shean, 2010), associated with 3D effects such as orographic precipitation. Models disagree about where ice should precipitate under different orbital conditions (Mischna et al., 2003; Levrard et al., 2004; Forget et al., 2006; Mischna and Richardson, 2006; Madeleine et al., 2009). This motivates follow-up GCM work.

Atmospheric collapse to form permanent CO_2 ice caps is more likely for Faint Young Sun insolation and for ~ 100 mbar initial P (Kahre et al., 2011) (Forget et al., 2012, submitted manuscript). However, snowmelt requires high ϕ , which is less favorable for atmospheric collapse. Will a CO_2 atmosphere that has collapsed at low ϕ reinflate on return to high ϕ ? A straightforward calculation suggests that atmospheres do not stay collapsed. Dividing a 100 mbar atmosphere by the current seasonal CO_2 exchange rate of ~ 3 mbar/yr gives a reinflation time of 30 yr, much shorter than orbital change timescales of 10^4 yr. Therefore the atmosphere is relatively unlikely to be collapsed for orbital conditions that optimize snowmelt.

Neglecting the lapse rate in surface temperature is a good approximation for current Mars, where surface temperature is set by radiative fluxes (Zalucha

et al., 2010). Results from the LMD GCM (Wordsworth et al. (2012); Forget et al., submitted manuscript) show that the adiabatic lapse rate is not large at 250 mbar but is important for $P_o \sim 500$ mbar. To cross-check, the Ames Mars GCM was run at 80 mbar for modern orbital conditions, topography, and luminosity. Only a weak increase in surface-temperature coupling to the adiabatic lapse rate was found relative to the 6 mbar case. Therefore, neglect of the adiabatic lapse rate coupling to surface temperature appears to be adequate for $P_o \sim 100$ mbar.

The model assumes that instantaneous values of e , ϕ and L_p are independent. Reverse integrations of the Solar System (obtained from <http://www.imcce.fr/Equipes/ASD/insola/mars/DATA/index.html>) show statistically significant correlation ($p < 0.0003$) between e and ϕ , but with a very small correlation coefficient ($|R| < 0.08$) and a sign that varies between integrations. The weakness of these correlations justifies treating each orbital parameter independently. Mean probabilities exceed median probabilities for high e , but the exceedance probability for $e = 0.15$ is ~ 0.8 over 4 Gya (Laskar, 2008).

We assume \mathbf{C} changes more slowly than \mathbf{O} , because post-Noachian rates of volcanic degassing, weathering, and loss to space are small compared to the atmospheric reservoir of CO_2 . This assumption does not consider volcanic- or impact-driven transients in ΔT .

We assume the freezing point depression for melting is not very large, which is appropriate for sulfates (e.g., $\Delta T \lesssim 4\text{K}$ for the magnesium sulfate - H_2O eutectic brine). Chloride brines allow liquid at much lower temperatures (Pollard et al., 1999; Fairén et al., 2009).

8.2 Comparison with other proposed mechanisms for sedimentary rock formation

Mechanisms for sedimentary rock formation on Mars must define sources of water, sediment, sulfur, and heat. In the ice-weathering model of Niles and Michalski (2009), the water source is an ice sheet. Sediment and sulfur is sourced from dust and gas trapped within the ice sheet. The heat source for weathering is the solid state greenhouse effect at shallow depths, and geothermal heating as the ice is buried. In the global-groundwater model (Andrews-Hanna et al., 2007, 2010; Andrews-Hanna and Lewis, 2011), strong greenhouse forcing warms the low latitudes to $> 273\text{K}$ (long term average). The water source is a deep, regional-to-global groundwater reservoir, which is recharged by precipitation or basal melting. Sulfur can be either from pyrite or from the atmosphere. The seasonal melting model implies conditions that are warmer and wetter than the ice-weathering model, but much colder and drier than the

1034 global-groundwater model. Snowmelt under a moderately thicker atmosphere
 1035 is the water source, and insolation under unusual orbital conditions supplies
 1036 heat. Sediment is atmospherically transported – ice nuclei, dust–storm de-
 1037 posits, saltating sand, ash, and fine-grained impact ejecta – and it is trapped
 1038 in the snowmelt area by aqueous cementation. The sulfur source is the atmo-
 1039 sphere.

1040 The main strength of the ice-weathering model is that it is (near-)uniformitarian.
 1041 Ice-sheet sulfate weathering is ongoing on Earth, and there is evidence for re-
 1042 cent sulfate formation on Mars (Mangold et al., 2010; Massé et al., 2012). Cur-
 1043 rent gaps in the ice-weathering model include the difficulty of explaining in-
 1044 terbedded runoff features (Grotzinger et al., 2006), except as post-sulfate re-
 1045 working, and a lack of a physical model for the proposed weathering mecha-
 1046 nism.

1047 Global groundwater models can explain the location of sedimentary rocks
 1048 and the diagenetic stratigraphy at Meridiani (Andrews-Hanna et al., 2007;
 1049 Andrews-Hanna and Lewis, 2011; Hurowitz et al., 2010). The global ground-
 1050 water model is internally self-consistent and complete. Upwelling rates are
 1051 consistent with inferred sediment accumulation rates. The discovery of gyp-
 1052 sum veins in material eroded from the Shoemaker Formation ejecta in Endeav-
 1053 our Crater has been interpreted as evidence for bottom-up groundwater flow
 1054 (Squyres and Athena Science Team, 2012). Chaos terrain strongly suggests
 1055 Mars had cooled enough to form a cryosphere that could modulate ground-
 1056 water release. Therefore, even in the groundwater model, post-chaos interior
 1057 layered deposits must have formed via a mechanism consistent with $\bar{T} < 273$,
 1058 such as spring flow (Pollard et al., 1999; Grasby et al., 2003).

1059 The advantages of the snowmelt model over previous models for the sedimentary-
 1060 rock water source are as follows. The snowmelt model arises from a self-
 1061 consistent climate solution (§4 – §5), liquid water production can “start and
 1062 stop” rapidly relative to Milankovitch cycles, and the equatorial concentra-
 1063 tion of sedimentary rocks emerges naturally (§5). The snowmelt model can
 1064 account for the global distribution of sedimentary rocks (§6). In the snowmelt
 1065 model, the sedimentary rocks form more or less in their current locations,
 1066 with their current layer orientations, and in their current shapes. Most sedi-
 1067 mentary rocks are now in moat-bounded mounds, filling craters and canyons.
 1068 Groundwater models imply removal of $\gg 10^6$ km³ of siliciclastic rock to an
 1069 unknown sink (Zabrusky et al., 2012; Andrews-Hanna, 2012). This removal
 1070 is mediated by a major phase of aeolian erosion which produces the moats.
 1071 Structural deformation is also required to tilt the near-horizontal primary dips
 1072 expected for playa-like deposition to the observed present-day draping dips.
 1073 There is no need to appeal to large-scale postdepositional modification in ei-
 1074 ther the snowmelt model or the ice-weathering model. Notwithstanding these
 1075 advantages, the snowmelt model assumes that precipitation is uniform, but

1076 in reality it must have been spatially variable. The snowmelt model also does
1077 not include a physical model for any of the steps linking melt generation to
1078 bedrock formation.

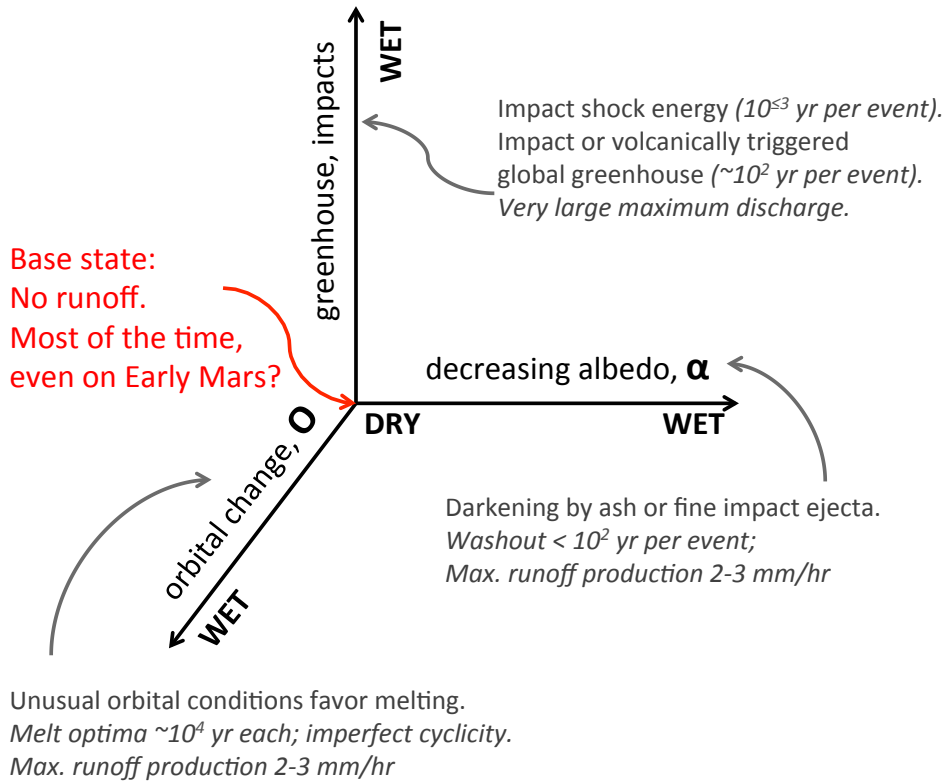
1079 8.3 *Atmospheric evolution and the decline of sedimentary rock formation*

1080 Few sedimentary rocks form on Mars now, and there is minimal surface liquid
1081 water. The only evidence for surface liquid water at the *Opportunity* land-
1082 ing site since the current deflation surface was established is minor Na/Cl-
1083 enriched veneers and rinds (Knoll et al., 2008). The simplest explanation for
1084 these changes is CO₂ escape to space. The 2013 MAVEN mission will constrain
1085 the present-day rate of escape to space. Supposing a 50-150 mbar atmosphere
1086 at sedimentary-rock time (Figure 15, marginally consistent with Manga et al.
1087 (2012)), a modern reservoir of 12 mbar (Phillips et al., 2011), and that soil
1088 carbonate formation has been unimportant, a loss to space of ~40-140 mbar
1089 over 3.5 Gya is predicted. Total loss of ~40-140 mbar is higher than previ-
1090 ous estimates of 0.8-43 mbar over 3.5 Gya from extrapolation of ASPERA-3
1091 measurements (Barabash et al., 2007), and 2.6-21.5 mbar from fits to MHD
1092 models by Manning et al. (2011). An alternative loss mechanism for CO₂ is
1093 uptake by carbonate weathering (Kahn, 1985; Manning et al., 2006; Boynton
1094 et al., 2009; Kite et al., 2011a). However, many sedimentary rocks contain sul-
1095 fates, and small amounts of SO₂ prevent carbonate precipitation (Bullock and
1096 Moore, 2007; Halevy and Schrag, 2009). Another alternative is that orbital
1097 conditions needed to drive melting were sampled early in Mars history, but
1098 not subsequently.

1099 8.4 *The Early Mars climate trade space*

1100 This paper has emphasized unusual orbital conditions, but more than one
1101 mechanism could lead to melting of snow or ice on Mars (Figure 18).

1102 Deposition of ash or fine-grained impact ejecta can lower albedo, driving tran-
1103 sient runoff events on Early Mars (Equation 4). Unweathered silicates provide
1104 the trigger for their own alteration by darkening the snow. In addition to
1105 Gale Crater (§7.1), this α -hypothesis is relevant for phyllosilicate formation
1106 at Mawrth: a regionally extensive, layered deposit which may be consistent
1107 with top-down alteration of ash (Noe Dobrea et al., 2010; McKeown et al.,
1108 2009; Bishop et al., 2008; Wray et al., 2008; Michalski and Noe Dobrea, 2007;
1109 Loizeau et al., 2007). Layered clays generally predate the sulfate rocks, consis-
1110 tent with decline of volcanism and impacts in the Early Hesperian (Ehlmann
1111 et al., 2011). Albedo effects are the primary regulator of spatial and temporal



Alternative: Early Mars base state is wet.
Test: Predicts $>10^5$ yr of continuous wet conditions.

Fig. 18. The Early Mars climate trade space. Assuming runoff did not occur during most years on Early Mars, nonzero runoff can be produced by perturbing orbital conditions, reducing albedo, heating from greenhouse forcing or impact shock energy, or some combination (black axes). All mechanisms can produce runoff ~ 1 mm/hr, but are distinguishable (gray arrows) by their limiting runoff and by their timescale.

1112 (Hall et al., 2010) melt production in the Antarctic Dry Valleys. Experimentally
1113 dusted ($13\text{--}100\text{ g/m}^2$ fine sand) patches of an Antarctic Dry Valleys
1114 glacier surface showed sharp increases in melt rate (Lewis, 2001). Antarctic
1115 Dry Valleys snowpack melts faster when it is buried beneath sand (Heldmann,
1116 2012), provided the sand cover is mm-dm thick.

1117 Heating from longwave forcing or conduction (ΔT or impact ejecta heating)
1118 could also drive melting. Increased $LW\downarrow$ could result from clouds (but see
1119 Colaprete and Toon (2003)) or short-lived pulses of volcanogenic gases (Halevy
1120 et al., 2007; Johnson et al., 2008; Tian et al., 2010; Halevy and Head, 2012)).
1121 Melting by impact ejecta is discussed by Mangold et al. (2012) and Noe Dobrea
1122 et al. (2010).

1123 Solar luminosity is the final dimension of the Early Mars climate trade space.

1124 This paper uses a standard solar model (Bahcall et al., 2001) that is consistent
 1125 with solar neutrinos and helioseismology, but not elemental abundances in
 1126 the photosphere (Asplund et al., 2005). Enhanced mass loss from the young
 1127 Sun would help resolve this discrepancy, and would make the young Sun more
 1128 luminous (Guzik and Mussack, 2010; Turck-Chièze et al., 2011). To change the
 1129 conclusion that the Sun was faint at the time the sedimentary rocks formed,
 1130 the Sun’s subsequent mass loss rate must have been 2 orders of magnitude
 1131 higher than inferred from nearby solar-analog stars (Wood et al., 2005; Minton
 1132 and Malhotra, 2007).

1133 Future work could determine which mechanism is responsible using geologic
 1134 observations that constrain discharge and timescale. Albedo reduction events
 1135 are short-lived, with runoff production that cannot exceed 2-3 mm/hr. Opti-
 1136 mal orbital conditions are relatively long-lived, but again runoff production is
 1137 limited by sunlight energy and cannot exceed 2-3 mm/hr. Volcanic- or impact-
 1138 driven events are short-lived, but with potentially very large discharge.

1139 The climates considered in this paper are extremely cold, comparable in melt
 1140 production to the coast of Antarctica (Liston and Winther, 2005). These
 1141 climates can produce enough water for aqueous alteration, but struggle to
 1142 match peak-runoff constraints. Therefore, there is scope for exploring warmer
 1143 snowmelt-producing climates, comparable to the coast of Greenland. One pos-
 1144 sibility is that cementation of the sedimentary rocks is the result of opti-
 1145 mal orbital conditions, but that river deposits interbedded with those rocks
 1146 (Williams, 2007) record additional transient (non-orbital) warming.

1147 **9 Summary and conclusions**

1148 The work presented here has two parts. First, a seasonal melting framework
 1149 for Mars has been developed that relates candidate paleoclimate parameters
 1150 to the production of seasonal meltwater. This framework has potentially broad
 1151 applications. Second, the model has been applied to a specific problem: the
 1152 origin and distribution of sedimentary rocks on Mars.

1153 Seasonal melting on Mars is the product of tides of light and tides of ice,
 1154 which move around the planet on Milankovitch frequencies. The peaks of
 1155 these tides rarely intersect. When they do, melting occurs. This water source
 1156 may contribute to sedimentary rock formation. The main conclusions from
 1157 this work are as follows:-

- 1158 • Order-of-magnitude calculations indicate that snowmelt is a sufficient water
- 1159 source for sedimentary rock formation on a cold Early Mars.
- 1160 • The distribution of sedimentary rocks on Mars is narrowly concentrated at

- 1161 equatorial latitudes and at low elevations.
- 1162 • The optimal spin-orbital conditions for snowmelt in cold traps on Mars are
1163 high obliquity, longitude of perihelion aligned with equinox, and eccentric-
1164 ity as high as possible. Melting then occurs in the early afternoon, at the
1165 equator, during perihelion equinox season.
- 1166 • A model of snowmelt on Early Mars has been presented, which uses a
1167 potential-well approximation to track cold traps for all orbital conditions.
1168 Integrated over all orbital conditions on an idealized flat planet, and assum-
1169 ing snowpack with the albedo of dust and a ~ 100 mbar pure CO_2 atmo-
1170 sphere, the model predicts a narrow equatorial concentration of snowmelt is
1171 predicted if warm-season snow is tightly confined to cold traps. A broad low-
1172 latitude belt of snowmelt is predicted if warm-season snow is more broadly
1173 dispersed.
- 1174 • With MOLA topography, atmospheric pressure > 100 mbar drives snow
1175 to high ground. High f_{snow} allows snowmelt on low ground even at high
1176 P . Sedimentary rocks are not on high ground, so either f_{snow} was high,
1177 snowmelt was not the water source for the sedimentary rocks, or P was
1178 $\lesssim 100$ mbar at time of sedimentary rocks.
- 1179 • With MOLA topography, a large swathe of parameter space produces a
1180 snowmelt distribution that is a good match to sedimentary rock locations.
1181 Enough water is produced to satisfy mass balance for aqueous alteration of
1182 sedimentary rock.
- 1183 • Early Mars climate states that produce the best fit to the spatial distribution
1184 of sedimentary rocks on Mars are cold. Much warmer climates would lead to
1185 snowmelt over a large swath of the planet, inconsistent with observations.
- 1186 • Climates that permit surface liquid water on Mars usually predict snowmelt
1187 at Gale Crater. Therefore, if MSL does not find evidence for a snowmelt
1188 contribution to sedimentary rock formation at Gale Crater, this would be a
1189 decisive failure of the model presented here.
- 1190 • Specific predictions for MSL at Gale Crater include generally homogenous
1191 aqueous chemical processing on ascending the mound, with clay layers cor-
1192 responding to a change in siliciclastic input, rather than a change in global
1193 environmental chemistry. The Gale Crater mound should have experienced
1194 wet/dry orbital cycles, with wet events only during optimal conditions. Ev-
1195 idence for vertical fluid flow over distances comparable to the height of the
1196 Gale Crater mound would be a major failure of the model presented here.
- 1197 • This is the first physical model to identify Gale Crater as a hemispheric
1198 maximum for sedimentary rock formation on Mars. The model therefore has
1199 the potential to relate observations at Gale Crater to global habitability.

A Data analysis

Locations in the MOC NA sedimentary rock database are likely to be strongly correlated with the true distribution of sedimentary rocks on Mars, even though MOC NA did not sample the planet uniformly. MOC NA took 97,000 images of Mars; 4% showed sedimentary rocks (Malin et al., 2010) (<http://marsjournal.org/contents/2010/0001/files/figure16.txt>). Although MOC NA imaged only 5.5% of Mars' surface (Malin et al., 2010), the Mars Reconnaissance Orbiter Context Camera (CTX) has surveyed >75% of the planet at comparable resolution to MOC NA (April 2012 Malin Space Science Systems press release, <http://www.msss.com/news/index.php?id=43>) and has not found large areas of sedimentary rock missed by MOC NA. MOC NA targets were selected on a 1-month rolling cycle on the basis of Viking imagery, previous MOC images, and the demands of other Mars missions (Malin et al., 2010). Sedimentary rocks were among the highest scientific priorities of the MOC NA investigation (Malin and Edgett, 2000; Malin et al., 2010). In the same way that oil wells are drilled more frequently in productive basins, there is a high density of MOC NA images in areas of sedimentary rocks identified early in the mission. Maps of the relative abundance of sedimentary rocks show only minor changes when defined using the fraction of MOC NA images showing sedimentary rocks within a given spatial bin instead of the absolute number of sedimentary rock observations in a given spatial bin.

The definition of sedimentary rock used by the MOC NA team excludes at least two areas that are sedimentary in origin, the Terra Sirenum drape deposit and a large part of the Medusae Fossae Formation (Grant et al., 2010; Bradley et al., 2002). However, the Terra Sirenum drape deposit has a distinct phyllosilicate-rich mineralogy from the sulphate-bearing sedimentary rocks that are the focus of this paper (Ehlmann et al., 2011), and the entire area of the Medusae Fossae Formation is consistently predicted to be a global near-maximum in sedimentary rock accumulation by our orbitally-integrated model output. Therefore, neither of these omissions from the database is important to the data-model comparison.

Because formation of sulfate-bearing sedimentary rocks peaked in the Hesperian, Terminal Hesperian and Amazonian terrain may conceal underlying sedimentary rocks and should be excluded from the analysis. The currently available global geological map of Mars (Skinner et al., 2006) is a digital renovation of Viking-era hardcopy maps (Scott and Tanaka, 1986; Greeley and Guest, 1987; Tanaka and Scott, 1987). Instead of using the old map, the edge of young materials was traced using the USGS Mars Global GIS as a base. The resulting ("K12") mask covers 45% of the planet, but only 3.5% of the images of sedimentary rocks ($n = 105$). These 3.5% are mostly from the Medusae Fossae Formation and the plateaux surrounding Valles Marineris. These rocks

1241 appear to represent a late tail in sedimentary rock formation, and so are re-
 1242 tained in the plots of elevation and latitude dependence (Figure 2). Omitting
 1243 them does not change our conclusions. To check that our results do not de-
 1244 pend on the idiosyncrasies of the tracing, Figure 2 from Nimmo and Tanaka
 1245 (2005) was georeferenced onto MOLA topography and areas that they mapped
 1246 as Late Hesperian or Amazonian materials (34% of the planet) were traced
 1247 (the “NT05 map”). K12 draws the dichotomy boundary close to the highland
 1248 break-in slope, whereas NT05 draws the boundary near the lowland edge of
 1249 the fretted terrain. Unlike NT05, K12 includes the Medusae Fossae Forma-
 1250 tion as potentially dating from the sedimentary rock era, in virtue of recent
 1251 results proving a Hesperian age for large parts of the formation ((Kerber and
 1252 Head, 2010; Zimbelman and Scheidt, 2012)). The elevation and latitude re-
 1253 sults shown in Figure 2 show little change between NT05 and K12 masks.
 1254 Earlier versions of the analysis presented in this paper (Kite et al., 2011b,c)
 1255 used NT05, and reached unchanged conclusions. NT05 reflects the modern
 1256 understanding of Mars geology better than Skinner et al. (2006), but it was
 1257 intended as a low resolution overview. We believe that K12 is better suited
 1258 than either the Viking-era maps or NT05 for the purpose of masking out
 1259 post-sedimentary-rock terrain.

1260 Image-center coordinates are assumed to be close to the locations of sedimen-
 1261 tary rocks. Image footprints can be large for orbits early in the MOC NA
 1262 mission, so these phases are excluded.

1263 The Valles Marineris are a unique tectonic feature containing many sedimen-
 1264 tary rocks. To make sure that conclusions are insensitive to this unique tectonic
 1265 feature, all data was excluded within a large “Valles Marineris box” (260E -
 1266 330E, 20S - 20N). This did not significantly change the latitude or elevation
 1267 dependence.

1268 The equatorial concentration of sedimentary rocks was previously noted in
 1269 uncorrected data by Lewis (2009) (unpublished PhD thesis).

1270 **B Details of thermal model**

1271 *Radiative terms:* A line-by-line radiative transfer model of the atmosphere
 1272 (Halevy et al., 2009) is used to populate two look-up tables:– $LW\downarrow$ as a func-
 1273 tion of T_1 and P ; and $SW\downarrow$ as a function of P and solar zenith angle. The
 1274 radiative transfer model, which for simplicity assumes a clear-sky, pure CO_2
 1275 atmosphere with no clouds or dust, is not run to radiative-convective equilib-
 1276 rium. Instead, for each combination of surface P , T , α , and solar zenith angle,
 1277 an atmospheric P - T structure is prescribed and the resulting radiative fluxes
 1278 are calculated. Following the approach of Kasting (1991), the tropospheric

lapse rate is dry adiabatic and the stratosphere is approximated as isothermal with a temperature of 167 K. A two-stream approximation to the equations of diffuse radiative transfer (which accounts for multiple scattering) is solved over a wavelength grid with a spectral resolution of 1 cm^{-1} at frequencies lower than $10,000 \text{ cm}^{-1}$ and a spectral resolution of 10 cm^{-1} at higher frequencies. The error induced by this spectral resolution relative to high resolution calculations is small compared to the uncertainties in the other model parameters (Halevy et al., 2009). The parameterisation of collision-induced absorption is the same as in Wordsworth et al. (2010), and is based on measurements by Baranov et al. (2004) and calculations by Gruszka and Borysow (1997, 1998).

The atmospheric temperature profile corresponding to $LW\downarrow$ is pinned to the diurnal average T_1 . Mars' bulk atmospheric radiative relaxation time is ~ 2 days at 6 mbar surface pressure (Goody and Belton, 1967; Eckermann et al., 2011), and increases in proportion to atmospheric density. It is assumed to be large for the P relevant to melting (>50 mbar).

Free convective terms: The turbulent flux parameterizations closely follow Dundas and Byrne (2010). Sensible heat loss by free convection is:

$$S_{fr} = 0.14(T - T_a)k_a \left(\left(\frac{C_p \nu_a \rho_a}{k_a} \right) \left(\frac{g}{\nu_a^2} \right) \left(\frac{\Delta \rho}{\rho_a} \right) \right)^{1/3} \quad (1)$$

where T_a is the atmospheric temperature, k_a is the atmospheric thermal conductivity, C_p is specific heat capacity of air, ν_a is viscosity of air, ρ_a is density of air, g is Mars gravity, and $\Delta \rho / \rho_a$ is the difference in density between air in equilibrium with the ground and air overlying the surface layer. $\Delta \rho / \rho_a$ is given by

$$\frac{\Delta \rho}{\rho} = \frac{(m_c - m_w)e_{sat}(1 - r_h)}{m_c P} \quad (2)$$

Here, m_c is the molar mass of CO_2 , m_w is the molar mass of H_2O , r_h is the relative humidity of the overlying atmosphere, and e_{sat} is the saturation vapor pressure over water ice. The expression for $\Delta \rho$ assumes that water vapor is a minor atmospheric constituent.

T_a is parameterized as (Dundas and Byrne, 2010)

$$T_a = T_{min}^{b_{DB}} T^{1-b_{DB}} \quad (3)$$

where T_{min} is the coldest (nighttime) surface temperature experienced by the model, and b_{DB} is the Dundas-Byrne 'b', a fitting parameter. This is an empir-

1308 ical model motivated by Viking 2 measurements (Dundas and Byrne, 2010).
 1309 b_{DB} decreases as P increases, because atmosphere-surface turbulent coupling
 1310 strengthens. $b_{DB}(P)$ is obtained by fitting to the output of GCM runs at 7,
 1311 50, and 80 mbar which employed a version of the NASA Ames Mars GCM
 1312 described in Haberle et al. (1993) and Kahre et al. (2006). Specifically, $b_{DB}(P)$
 1313 is fit to the global and annual average of the temperature difference between
 1314 the surface and the near-surface atmosphere for local times from 11:00-13:00.

1315 We let

$$L_{fr} = L_e 0.14 \Delta \eta \rho_a D_a \left(\left(\frac{\nu_a}{D_a} \right) \left(\frac{g}{\nu_a^2} \right) \left(\frac{\Delta \rho}{\rho} \right) \right)^{1/3} \quad (4)$$

1316 where L_e is the latent heat of evaporation, $\Delta \eta$ is the difference between at-
 1317 mosphere and surface water mass fractions, and D_a is the diffusion coefficient
 1318 of H_2O in CO_2 .

1319 *Forced convective terms:* Sensible heat lost by forced convection is given by:

$$S_{fo} = \rho_a C_p u_s A (T_a - T) \quad (5)$$

1320 where u_s is the near-surface wind speed. Near-surface winds are controlled
 1321 by planetary boundary layer turbulence which serves to mix the atmosphere
 1322 vertically, so $S_{fo} \neq 0$ is consistent with the assumption of no meridional heat
 1323 transport. The drag coefficient A is given by

$$A = \left(\frac{A_{vonk}^2}{\ln(z_{anem}/z_o)^2} \right) \quad (6)$$

1324 where A_{vonk} is von Karman's constant, z_{anem} is anemometer height, and z_o is
 1325 surface roughness.

1326 Near-surface wind speed u_s in the NASA Ames Mars GCM decreases with
 1327 increasing P and decreasing solar luminosity. The four-season average of Eu-
 1328 ropean Mars Climate Database ("MY24" simulation) globally-averaged near-
 1329 surface wind speeds at the present epoch is 3.37 m/s (Millour et al., 2008).
 1330 This is extrapolated for $P \leq 290$ mbar using a logarithmic dependence of u_s on
 1331 P fitted to the global and annual average of Ames Mars GCM model surface
 1332 wind speed for initial pressures of 7, 50 and 80 mbar. u_s is lowered by a factor
 1333 of 1.08 for the Faint Young Sun using the ratio of wind speeds for two 50 mbar
 1334 Ames Mars GCM simulations that differ only in solar luminosity. Simulations
 1335 suggest u_s increases with ϕ (Haberle et al., 2003), but this is ignored. Figure
 1336 9 shows the sensitivity of results to $u_s = f(P)$ and $u_s \neq f(P)$.

1337 Latent heat losses by forced convection are given by:

$$L_{fo} = L_e \frac{M_w}{kT_{bl}} u_s (e_{sat}(1 - r_h)) \quad (7)$$

1338 where M_w is the molecular mass of water, and k is Boltzmann's constant.
1339 Latent heat fluxes for dirty snow are calculated assuming that the entire ex-
1340 posed surface area is water ice. Dirt concentrations are at the percent level by
1341 volume, or less, for all results presented here, so this is acceptable.

1342 The free and forced fluxes are summed together, rather than considering only
1343 the dominant term. This matches the functional form of Mars-chamber data
1344 (Chittenden et al., 2008) and is the standard approach in Mars research (Dun-
1345 das and Byrne, 2010; Williams et al., 2008; Toon et al., 1980). However, sum-
1346 ming the terms is an idealization that may overestimate cooling.

1347 *Melt handling:* Melt occurs when $T_K > (273.15\text{K} - \Delta T)$. ΔT is a freezing-point
1348 depression. It can also be interpreted as any non-CO₂ warming due to water
1349 vapor, ice clouds, or SO₂, stochastic fluctuations in material properties around
1350 those assumed in Table 1, or a higher solar luminosity. Additional greenhouse
1351 warming (freezing point at 273.15K) implies greater turbulent and $LW\uparrow$ losses
1352 at melting than freezing-point depression (freezing point at $273.15\text{K} - \Delta T$),
1353 but ΔT is small so this difference is ignored here.

1354 Total melt present and total melt produced are tracked during the sol. Melt is
1355 not permitted to drain, and the melt fraction is not allowed to affect snowpack
1356 material properties except to buffer temperature during refreezing (Liston and
1357 Winther, 2005).

1358 Ablation of the snowpack surface by sublimation is not directly tracked. The
1359 effect on sublimation on snowpack survival is treated indirectly, through the
1360 potential-well approximation (§4.3). However, ablation also affects snowpack
1361 temperature. Implied sublimation rates are ~ 0.5 mm/sol for conditions favor-
1362 able to melting. Movement of the snow surface down into the cold snowpack
1363 corresponds to advection of cold snow upwards (relative to the surface). Snow-
1364 pack thermal diffusivity is $\sim 2 \times 10^{-7}$ m²/s. Melting at depths greater than
1365 $\sim \kappa/u_{subl} \sim 4$ cm may be suppressed by this advective effect.

1366 *Run conditions.* Conductive cooling is found by matrix inversion. Vertical res-
1367 olution is ≈ 2.5 mm for nominal parameters, which is $0.033\times$ the analytic di-
1368 urnal skin depth. Time resolution is 12s, and the lower boundary condition is
1369 insulating.

1370 The initial condition at the surface is slightly cooler than radiative equilib-
1371 rium, decaying to the energy-weighted diurnal average temperature at depth

with an e-folding depth equal to the diurnal skin depth. The model is integrated forwards in time for several sols using constant seasonal forcing until the maximum T_1 on successive sols has converged (to $<0.01\text{K}$) and the diurnal-maximum melt column (if any) has converged to $<0.018\text{ kg/m}^2$. For polar summers, convergence can take an extremely long time as the melt zone spreads to cover the entire snowpack, so the integration stops after ~ 8 sols even if the convergence criteria are not met.

There is no meridional heat transport, seasonal thermal inertia, or CO_2 cycle. Temperatures are not allowed to fall below the CO_2 condensation point. For each spatial location, the model is run for many seasons (L_s). The converged output is then interpolated on a grid equally spaced in time to recover annual means.

Details about melt-likelihood map construction. Results in this paper are based on grids of runs at $\phi = \{0^\circ, 10^\circ, 20^\circ, \dots, 80^\circ\}$, $e = \{0, 0.03, 0.06, 0.09, 0.115, 0.13, 0.145, 0.16\}$, $L_p = \{0^\circ, 15^\circ, 30^\circ, \dots, 90^\circ\}$ (with mirroring to build up a full precession cycle), $L_s = \{0^\circ, 22.5^\circ, 45^\circ, \dots, 337.5^\circ\}$, and latitude $\{-90^\circ, -80^\circ, \dots, 90^\circ\}$, giving 1.5×10^6 snowpack thermal model runs for each **C**. Quoted results at intermediate values result from interpolation. Statements about **C** are based on interpolation in a grid of runs at $P = \{4, 8, 16, 24, 48\} \times 610\text{ Pa} \equiv \{24, 49, 98, 146, 293\}\text{ mbar}$, with $\Delta T = 0\text{K}$. ΔT and f_{snow} were varied in postprocessing.

To remove longitudinal stripes of high snow probability in the Northern Plains that are artifacts of finite model resolution in \mathbf{O}' and latitude, the step function in $(f_{\text{snow}} - f)$ is replaced by a linear ramp in $(f_{\text{snow}} - f)$. This is a minor adjustment.

C Snowpack radiative transfer

Crystalline water ice is opaque in the thermal infrared, but almost transparent to visible light. The resulting solid-state greenhouse effect enhances snowmelt (Clow, 1987; Brandt and Warren, 1993). The purpose of the solid-state greenhouse parameterization in this paper is to self-consistently model the tradeoff between snowpack broadband albedo (α) and subsurface absorption of sunlight. This does not require precisely calculating α as a function of dust content, so the model uses simple linear approximations to the radiative transfer equations developed for widely-separated atmospheric aerosols (e.g. Kieffer (1990); Calvin et al. (2009)). Although more sophisticated models can be employed to take account of aspherical particles, near-field effects, and heterogeneous compositions (e.g. Cull et al., 2010; Yang et al., 2002), the lack of consensus on their importance leads us to not include them in our algorithm.

The solid-state greenhouse parameterization uses the snow radiative transfer model of Brandt and Warren (1993). Ice refractive indices are from Warren and Brandt (2008), and are converted to Henyey-Greenstein parameters using a standard Mie code following Bohren and Huffman (1983). Mars dust optical parameters are calculated using the refractive indices of Wolff et al. (2006, 2009). An illustration of these parameters for the canonical atmospheric dust sizes is shown in Figure 1 of Madeleine et al. (2011), but we also employ larger sizes as well. The 2000 ASTM Standard Extraterrestrial Spectrum Reference E-490-00 is used to describe the wavelength dependence of the direct flux component; diffuse flux is neglected as being a minor perturbation. The young Sun was $\sim 100\text{K}$ cooler in the standard solar model. Solar reddening increases α by <0.01 , so the spectral shift is ignored here. The effect of small amounts of meltwater on α is minor (Warren, 1982) and is also ignored. The effects on wavelength-dependent direct-beam semi-infinite albedo (not shown) are broadly similar to the idealized “red dust” in Warren and Wiscombe (1980). Once optical properties are prescribed, the most important variables are dust content, effective dust grain radius, and effective ice grain radius. A given α can usually be obtained by several different combinations of these properties. The Brandt and Warren (1993) model is used to build a look-up table of fractional subsurface absorption as a function of these variables, plus direct-beam path length. This length is mapped to depth within soil by multiplying by the cosine of the zenith angle.

The radiative transfer model reproduces the trends found by Clow (1987). The larger values of the Martian dust single scattering albedo in the optical (Wolff and Clancy, 2003; Wolff et al., 2006, 2009) reduce the amount of melting for a given dust concentration. Ice grain size growth is slow in Mars’ present day polar caps (Kieffer, 1990) but much faster under the near-melting conditions that are important for the model presented here. We adopt an effective size of 1 mm, corresponding to observed ice-grain radii in hoar layers in Earth snowpacks. Not surprisingly, there are no direct measurements of dust content in snow on Mars. Dust content in ice has been reported as “a few percent (up to at most around 30%)” by volume in the Northern Plains (Dundas and Byrne, 2010), and $\sim 15\%$ by mass in the SPLD (Zuber et al., 2007). We assume $\sim 2\%$ dust mass fraction by volume and a dust grain radius of $4\mu\text{m}$.

D Selection of snowpack material properties

Snow stability and peak temperatures are affected by material properties such as α and TI. Low (snowlike) TI (Carr and Head, 2003) is used here because snow precipitation is also predicted by all General Climate Models (GCMs) at high ϕ (e.g., Fastook et al. (2008); Mischna et al. (2003); Madeleine et al. (2009)). In addition, water ice precipitation was observed on Mars by the

1450 Phoenix lander (Whiteway et al. (2011), their Figure 1). High (icelike) TI
1451 suppresses the diurnal thermal wave and makes melting at the equator much
1452 more difficult.

1453 The impact of parts-per-thousand levels of dust on snowpack albedo and runoff
1454 is severe (Warren and Wiscombe, 1980; Warren, 1984). Present day observed
1455 and calculated Mars seasonal H₂O snow α is 0.25-0.4 (Vincendon et al., 2010;
1456 Kereszturi et al., 2011). α on the South Polar water ice cap is 0.30 (Titus et al.,
1457 2003). To allow melting to create gullies, snowpack surface-layer albedo must
1458 have been as low as 0.12 (Williams et al., 2009). Dust storms and dust devils
1459 occur every year, and caused major changes in regional and global albedo
1460 between 1978 and 2000 (Geissler, 2005) and between 2003 and 2007 (Putzig
1461 and Mellon, 2007). Globe-encircling dust storms, which now occur every few
1462 years, are likely to occur twice every year at high ϕ (Haberle et al., 2003).
1463 Dust is required to supply ice nuclei for heterogenous nucleation. Therefore,
1464 it is reasonable to expect snowpack at high ϕ to be contaminated with dust.
1465 Given the likelihood of dust contamination, this paper assumes $\alpha = 0.28$, the
1466 same as Mars' light-toned dust continents. This requires a dust concentration
1467 of $O(1\%)$ (Appendix C). Other likely sources of darkening contaminants on
1468 Early Mars are volcanic ash and fine-grained impact ejecta. In the words of
1469 Warren (1984) "When snow melts, the impurities often tend to collect at the
1470 surface rather than washing away with the meltwater." Using a low α favors
1471 melting, which is conservative because the reconstructed paleoclimate will
1472 involve the smallest change from the current Mars climate that is consistent
1473 with the geological evidence.

1474 Figure 1 shows sensitivity to material properties. Increased TI damps the
1475 diurnal surface temperature cycle. Increasing albedo lowers surface tempera-
1476 ture for all times of day, especially near noon. The response of the maximum
1477 temperature within the snowpack (lower loop) is complicated by subsurface
1478 absorption of sunlight. This occurs at greater depths when dust concentration
1479 is decreased, even though the snowpack as a whole is more reflective. Because
1480 buried solar energy cannot easily escape, nighttime subsurface temperatures
1481 are increased by increasing the albedo. The location of maximum temperature
1482 moves to steadily greater depths during the night. The energy-burial effect is
1483 abruptly reversed shortly after dawn (18 hours after noon), when the location
1484 of maximum temperature returns to the near-surface.

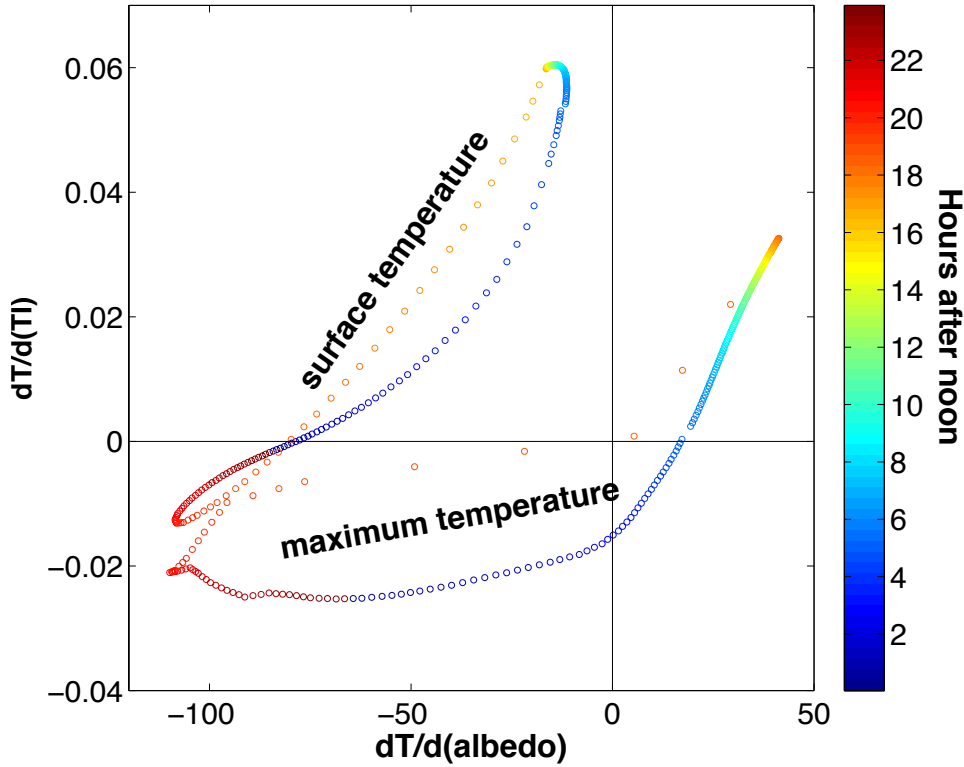


Fig. 1. Response of snowpack temperatures to adjusting material properties. Color corresponds to local hour angle (time after solar noon). The small jumps in the temperature loops are interpolation artifacts. $e=0.11$, $\phi=50^\circ$, and $L_p=0^\circ$.

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